UNIT 8

UPPER-AIR CHARTS AND THEIR ANALYSIS

FOREWORD

As important as the surface chart is in providing a picture of weather over an area, it is one dimensional. The surface chart, in conjunction with upper-air charts, provides us with the most complete picture possible of the three-dimensional distribution of wind and temperature in the atmosphere.

An accurate and complete analysis of all data (surface and upper air) does not guarantee a successful prognosis or forecast, but it definitely improves one's chances. As an analyst, you are much more than one who draws charts. You must be able to think, reason, and deduce from your products. Relating the weather aloft to that occurring at the surface is absolutely essential in understanding the processes working within the atmosphere. When you have concluded this unit, I hope you will have gained an understanding of upper-air charts and the insight required to become a successful analyst. This unit consists of five lessons. Lesson 1 covers upper-air analysis of constant-pressure charts and other types of supplementary analyses. Lesson 2 discusses the uses of constant-pressure charts. Lesson 3 discusses atmospheric circulation patterns and their appearance on upper-air charts. Lesson 4 relates the effects and importance of convergence and divergence, and Lesson 5 discusses the effects of rotational motion (vorticity) in the atmosphere.

UNIT 8—LESSON 1

UPPER-AIR ANALYSIS

OVERVIEW

As an AG 3, you have been exposed to upper-air charts and their analysis. Some AG 3s have plotted them, displayed them, and possibly even sketched contours on them. To further your understanding of these charts, we will discuss those elements which go into their makeup and analysis.

OUTLINE

Evaluation of data

Contour (isoheight) analysis

Isotherm analysis

Wind-temperature relationship

Frontal analysis

Isotach analysis

Jet stream analysis

Tropopause analysis

Supplementary upper-air analysis

UPPER-AIR ANALYSIS

The upper-air analyst is responsible for recognizing and resolving erroneous reports; extrapolating upper-air data from surface reports; drawing isopleths for various elements such as heights, temperatures, wind speeds, etc.; deriving upper-air information from satellite pictures; and relating the various levels analyzed.

Learning Objective: Determine the elements used in the preparation and analysis of upper-air charts.

EVALUATION OF DATA

Today's upper-air equipment is far better than that used only a few years ago, and it permits far more accurate readings of pressure, temperature, humidity, and wind. You must be aware, however, that the same type of errors and unrepresentative data that exists in surface reports also exists in upper-air reports. To a large extent these errors are attributable to the method used in determining the data. In many cases erroneous reports can be resolved into useable data. Common errors that sometimes can be corrected are communications and plotting errors, computation errors, and in the case of aircraft reports, erroneous position reports. You must exercise discretion in correcting such errors; however, do not throw away or disregard data unless you are absolutely sure it is in error and uncorrectable.

Data Priority

Observed data should be plotted according to the following priority:

1. Radiosonde and rawinsonde observations
2. Winds aloft observations by pilot balloons (pibals)

3. Weather reconnaissance aircraft observations (reccos)
4. Other aircraft observations (pireps)
5. Satellite-derived information

RADIOSONDES AND RAWINSONDES.— These are the most all-encompassing methods of gathering upper-air information and also the most accurate. Probable errors associated with sondes are listed in table 8-1-1.

WINDS ALOFT OBSERVATIONS (PIBALS).— The two biggest contributors to this data's being questionable are (1) the assumed-constant ascension rate of the balloon and (2) the timelag in the observer's reading of elevation and azimuth angles. These errors become magnified in areas of high winds or large vertical wind shears.

RECCOS AND PIREPS.— Flight altitude data is reasonably accurate and useful. Dropsonde data is considered somewhat less accurate than data gathered by balloon-borne sondes. Position errors make these observations stand out in analysis, because they fail to fit into the height pattern.

Table 8-1-1.—Probable Errors in Upper Air Data

Parameter	Probable error	Remarks
Raob temperatures	± 1°C to 400 mb ± 2°C above	
Raob heights 1,000 mb	± 6 meters per 305 meters of station elevation	Different temperature parameter used for reduction to 1,000 mb than to msl which results in inconsistencies of the order of 120 meters in extreme cases. In order to circumvent this, we directly convert the reported msl pressure to 1,000 mb.
850 mb	± 9 meters	
700 mb	± 12 meters	
500 mb	± 21 meters	
400 mb	± 30 meters	
300 mb	± 49 meters	
200 mb	± 70 meters	
150 mb	± 90 meters	
Thicknesses:		
1,000-700	± 12 meters	
1,000-500	± 21 meters	
500-300	± 27 meters	
Tropopause heights	± 10 mb	
Winds aloft directions	± 5 degrees	Due in part to rounding off to nearest 5 degrees in coding.
Winds aloft speeds	± 10 knots up to 500 mb for winds up to 50-75 kn	This error increases rapidly, especially for winds over 100 kt, usually due to low elevation angles of recording equipment. Also varies considerably with different types of equipment.

SATELLITES.— Satellite pictures of Earth allow us to verify the existence, intensity, and position of weather systems. In sparse-data areas, they are our primary means of completing an analysis.

Data Accuracy

Tests conducted to evaluate sonde accuracy have shown the data to be satisfactory if the observers have been fully trained in equipment operation and data evaluation.

TEMPERATURE.— Under controlled conditions, comparisons of sonde-recorded temperatures to those measured by thermometer show the sonde temperatures to be accurate within 1°C over 90% of the time.

PRESSURE.— The average sonde pressure error is from 2 to 4 mb. The absolute maximum error evaluated was 5 mb, and this occurred in only one percent of the sondes tested.

HUMIDITY.— This element is accurate to within 5% up to 700 mb, but maybe in error by as much as 20% at 300 mb. Because this element is recorded with respect to water (never ice) when a sonde passes through an ice crystal cloud layer the relative humidity computes at 70 to 80 percent. This is an accurate reading. A very rough estimate of humidities in below-freezing temperatures is obtained by decreasing the humidity 10% for every 10°C decrease in temperature.

UPPER WINDS.— The reliability of upper-wind data is dependent on the manner in which the data is obtained. Rawinsonde winds are for the most part very accurate. Winds obtained by radar tracking are the most accurate. See probable errors listed in table 8-1-1. This table is used by the National Weather Analysis Center.

Learning Objective: Identify the analysis procedures and use of thickness, time differential, jet stream, tropopause, and individual analyses.

CONTOUR (ISOHEIGHT) ANALYSIS

Today, few stations do manual analysis. Computers have for the most part replaced individual analysts. They are undoubtedly faster at deciphering reports, evaluating data for continuity and accuracy, and producing a myriad of products worldwide. However, there are times when computer generated products aren't available. At such times, you may be called upon to analyze upper level data. For this reason, and to better acquaint you with constant-pressure chart features, you should be familiar with analysis procedures. Just as there are recommended analysis procedures for the surface chart, there are similar procedures for constant-pressure charts. A recommended procedure is as follows:

1. Review past history.
2. Extrapolate heights in sparse data areas.
3. Sketch and label contours.
4. Sketch in troughs.
5. Evaluate slope of upper systems and their orientation. SYSTEMS MUST STACK.
6. Harden in troughs.
7. Harden in contours.
8. Label low and high height centers.

Past History

One basic consideration in the approach to all types of map analysis is that of history or continuity. Upper-level features do not change radically in short periods of time. Consequently, a valuable aid in contour analysis of any given pressure level is the past history of the contours at that level. The study of previous maps is essential, both as a key to the present situation and as a method of determining future movement and change in atmospheric systems.

Your first step in the analysis should be to check previous charts for accuracy, rationality, and any changes made to previous analyses based on information received after the analyses were completed. Fronts, troughs, and height centers do not normally appear and/or disappear in the 12 hours between analyses and any such occurrence should be viewed with suspicion.

The past position of all height centers should be entered on the current chart for at least 24 hours. These positions are normally entered in black ink with an X circumscribed with a circle and connected with a dashed line. The time and date are entered above the circle. The corrected positions of all troughs and ridges should be transposed onto the current chart in yellow pencil. Knowing the past positions of features is your first clue in locating their present positions. When an upper-level feature moves into an area containing few or no reports, you may have to extrapolate its movement, construct the upper-level field from surface data, and/or determine the new position from satellite pictures, if possible. Constructing the upper-level field from surface data is widely used in computer-generated products.

Extrapolating Upper-Level Heights

A common analysis practice is extrapolation of upper heights from sea-level reports. The scarcity of upper-air-reporting stations in many regions of the world and the absence of upper-level data from stations that do report require that this procedure be used. This is a common practice in computer-analyzed constant-pressure charts. Extrapolated heights are plotted enclosed in parentheses.

The thickness of a stratum or height of an upper level is computed using a known sea-level pressure and temperature and an assumed mean virtual temperature for the stratum. Tables and nomograms, derived from the hydrostatic equation, are also used for this purpose.

Figure 8-1-1 contains a nomogram for computing the height of the 700- or 500-mb level using a known surface temperature and an estimated upper-level temperature. The upper-level temperature is usually estimated from past analysis and compensated for any changes which may have occurred since the last analysis; i.e., any cold or warm air advection. A step-by-step extrapolation procedure follows:

1. Estimate the upper-level temperature.
2. Determine the height of the 1000-mb level.
3. Determine the thickness of the stratum, following nomogram directions.
4. Algebraically add the 1000-mb height and the thickness of the stratum obtained in step 3. This is the height of the upper level above sea level.

The 1000-mb height must be determined, because tables and nomograms use 1000 mb as the standard surface-pressure value. To compute its height, we assume 7 1/2 mb equals 60 meters of height, or 8 meters per millibar. Take the difference between the actual sea-level pressure (SLP) and 1000 mb and multiply this figure by 8. For example, a ship reports a sea-level pressure of 1015.0 mb. The difference between the reported value and 1000 mb is 15. Multiply this difference by 8. 15 x 8 = 120. The height of the 1000-mb level is 120 meters above sea level.

SLP(mb) – 1000 mb x 8 m/mb = 1000-mb height

Minus (–) values indicate 1000-mb heights below mean sea level, and plus (+) values indicate 1000-mb heights above mean sea level.

Given the following parameters, compute the height of the 700-mb level, using the nomogram shown in figure 8-1-1: Estimated 700-mb temperature – 5.0°C; sea-level temperature, 25.5°C. REMEMBER—the nomogram contains 700-mb heights and 500-mb heights. The 700-mb heights are to the left of the center line.

In step 4, algebraically add the thickness of the stratum just computed and the height of the 1000-mb level. For computational purposes, use the 120-meter height previously computed. The thickness of the 1000-700-mb stratum as computed from the above information should read approximately 2,957 meters. Because the reported sea-level pressure in this example is greater than the standard, you should have added the 120 meters to the height of the 1000-700-mb stratum. Your answer should read 3,077 meters. If the reported sea-level pressure is less than 1000 mb, subtract the 1000-mb height from the height of the stratum.

Since we are working with assumed or estimated temperatures and not those of an actual upper-air sounding, inversions anywhere in the stratum or non-representative temperatures at the surface or upper level will result in incorrect height computations. In the case of marked inversions of any type, the estimated height will be less than the "true" height, because the computed mean virtual temperature will be less (colder) than the true mean virtual temperature. Compensate for such inversions by using a higher estimate of the upper-level temperature. Higher estimates may be required in the following situations:

1. In the vicinity of high-pressure cells where subsidence inversions are present
2. Where surface inversions are indicated by stable weather phenomena such as fog
3. When a frontal surface exists below the level to be extrapolated

A study comparing the true thickness of the 1000-700-mb and 1000-500-mb stratums to extrapolated thicknesses using temperature averaging showed the extrapolated thicknesses

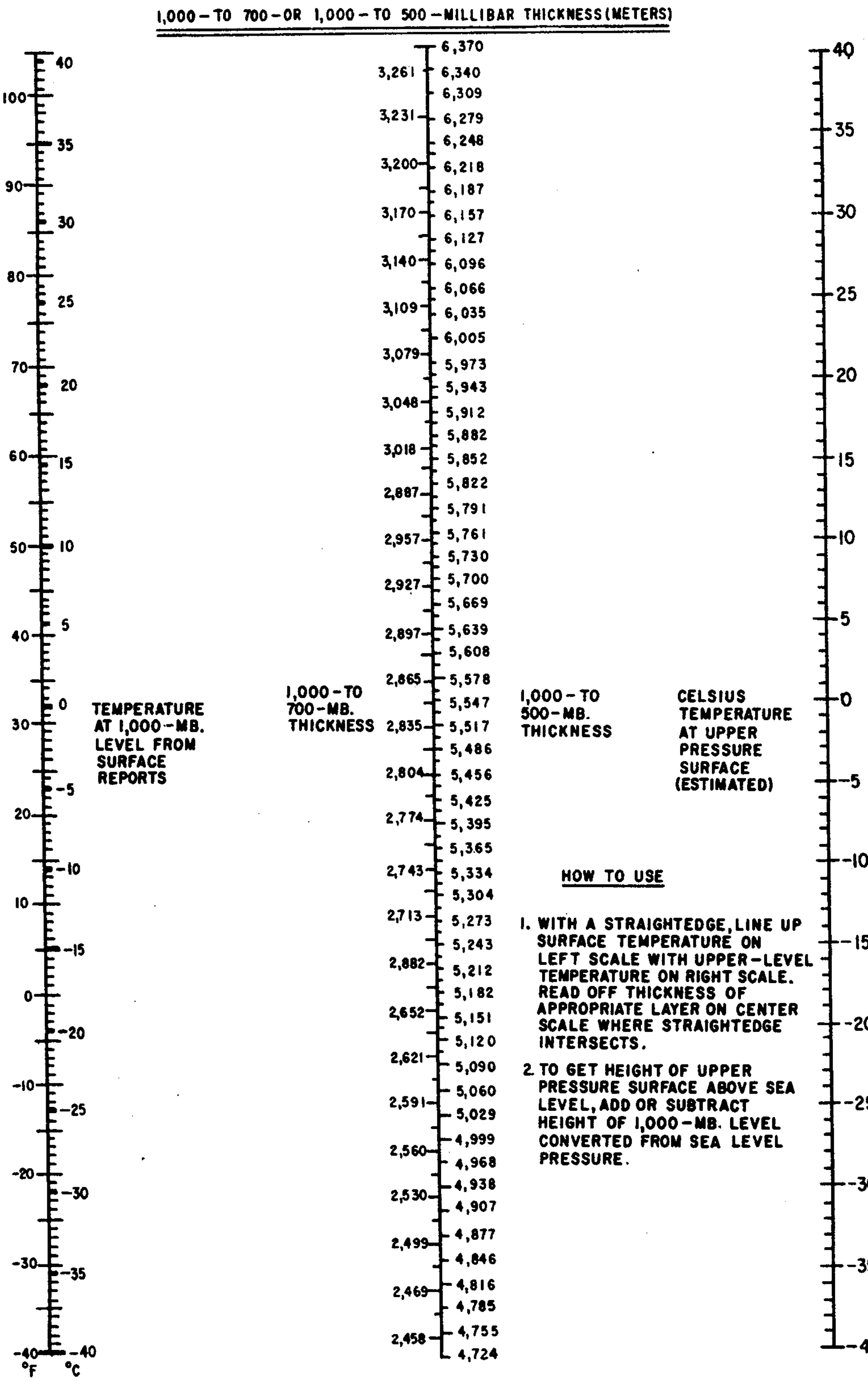

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Figure 8-1-1.—Nomograms for computing height of the 700-mb and 500-mb levels.

to be within 30 meters for the 1000-700-mb stratum and within 60 meters for the 1000-500-mb stratum in 90 percent of the cases conducted.

Sketching and Labeling Contours

Contours are the upper-air equivalent of isobars on surface charts. They connect points of equal height on constant-pressure charts. The primary contours (isoheights) are spaced at 30-, 60-, and 120-meter intervals. The 30-meter interval is used on charts below 500 mb; the 60-meter interval, on charts between 500 and 300 mb; and the 120-meter interval, on charts at 300 mb and above. Primary contours are represented by solid black lines. Some of the primary contours drawn on constant-pressure charts are as follows:

850 mb	700 mb	500 mb	300 mb	200 mb
1410 m	2910 m	5400 m	8760 m	11,400 m
1440	2940	5460	8880	11,520
1470	2970	5520	9000	11,640
*[1500	3000	5580	9120	11,760]
1530	3030	5640	9240	11,880
1560	3060	5700	9360	12,000
1590	3090	5760	9480	12,120

*The bracketed contours are closest to the standard heights

The above listing is not all inclusive; you may have to draw heights higher or lower than those listed.

Intermediate contours may be used when greater definition is needed. They are one-half the primary contour interval for the level in question. For example, the primary contours on the 500-mb chart are drawn for every 60 meters. The intermediate contours would be drawn at 30-meter intervals. Intermediate contours are represented by dashed black lines.

Contours should first be sketched lightly, using a black pencil, according to the reported heights and winds. Contours parallel the wind direction and are drawn following downwind, extrapolating between station height plots as necessary. The contours should conform to the wind direction. However, since most plotters do not plot wind directions with the aid of a wind rose or protractor, wind directions tend to be the least reliable of the plotted data. Be aware of the tens digit plotted at the end of the wind shaft. It is not uncommon to see the plotted wind direction differ from the reported value by as much as 20 degrees. For example, a 270° wind may get plotted anywhere from 250° to 290°. See figure 8-1-2.

While you are constructing the contour pattern, continually strive to accurately depict the wind field. Contour spacing is inversely proportional to wind speed. Only in areas of dense height reports where the winds are weaker (subgradient) or stronger (supergradient) than the gradient within which they appear is there any justification for drawing contours that do not conform to the reported wind. For stationary troughs and ridges, cyclonically curved contours are usually spaced more closely (tighter gradient) than anticyclonically curved contours of the same wind speed. Use geostrophic wind scales to aid in spacing contours in sparse-data areas.

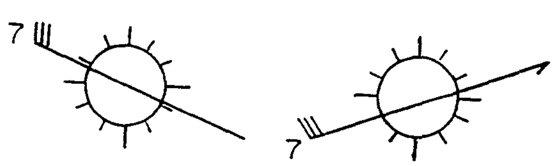

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Figure 8-1-2.—Example of an incorrectly plotted wind direction.

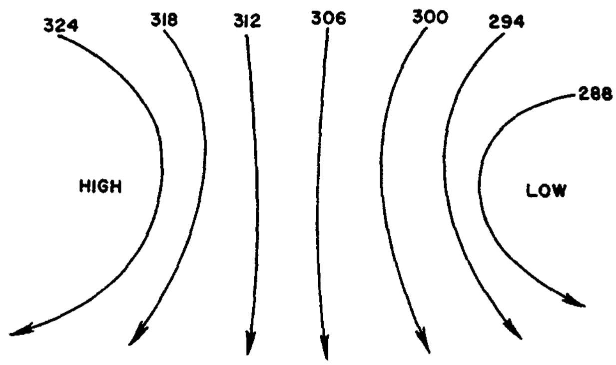

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Figure 8-1-3.—Contour pattern between a high and a low (700-mb heights in meters).

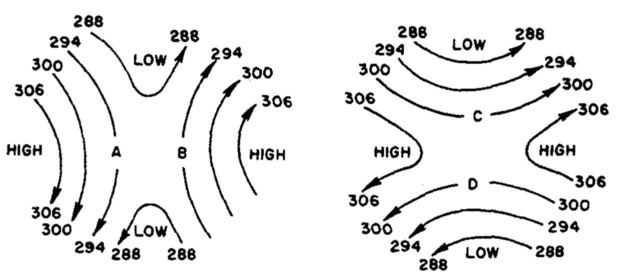

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Figure 8-1-4.—Contour pattern between adjacent highs and lows (700-mb heights in meters).

When you sketch preliminary contours, make smooth sweeping lines, bringing your entire arm into motion. Keep your eyes just ahead of the pencil. This will enable you to anticipate contour direction changes. Each contour will eventually form a loop or go off the chart. In no case do contours cross or touch contours of different values.

The following discussion of rules for drawing contours is illustrated by figures 8-1-3 and 8-1-4.

1. Two contours of the same value will never exist between adjacent highs and lows. See figure 8-1-3.

2. There will always be two contours with the same value between two adjacent highs and two adjacent lows. See figure 8-1-4.

3. Adjacent contours of the same value will always flow in opposite directions.

Some typical errors made by inexperienced analysts are illustrated in figure 8-1-5, to assist you in correctly drawing contours.

Contours are labeled in decimeters. Use the thousands, hundreds, and tens digits. For example, the 3060-meter contour on the 700-mb chart is labeled 306. The units digit is dropped. On the

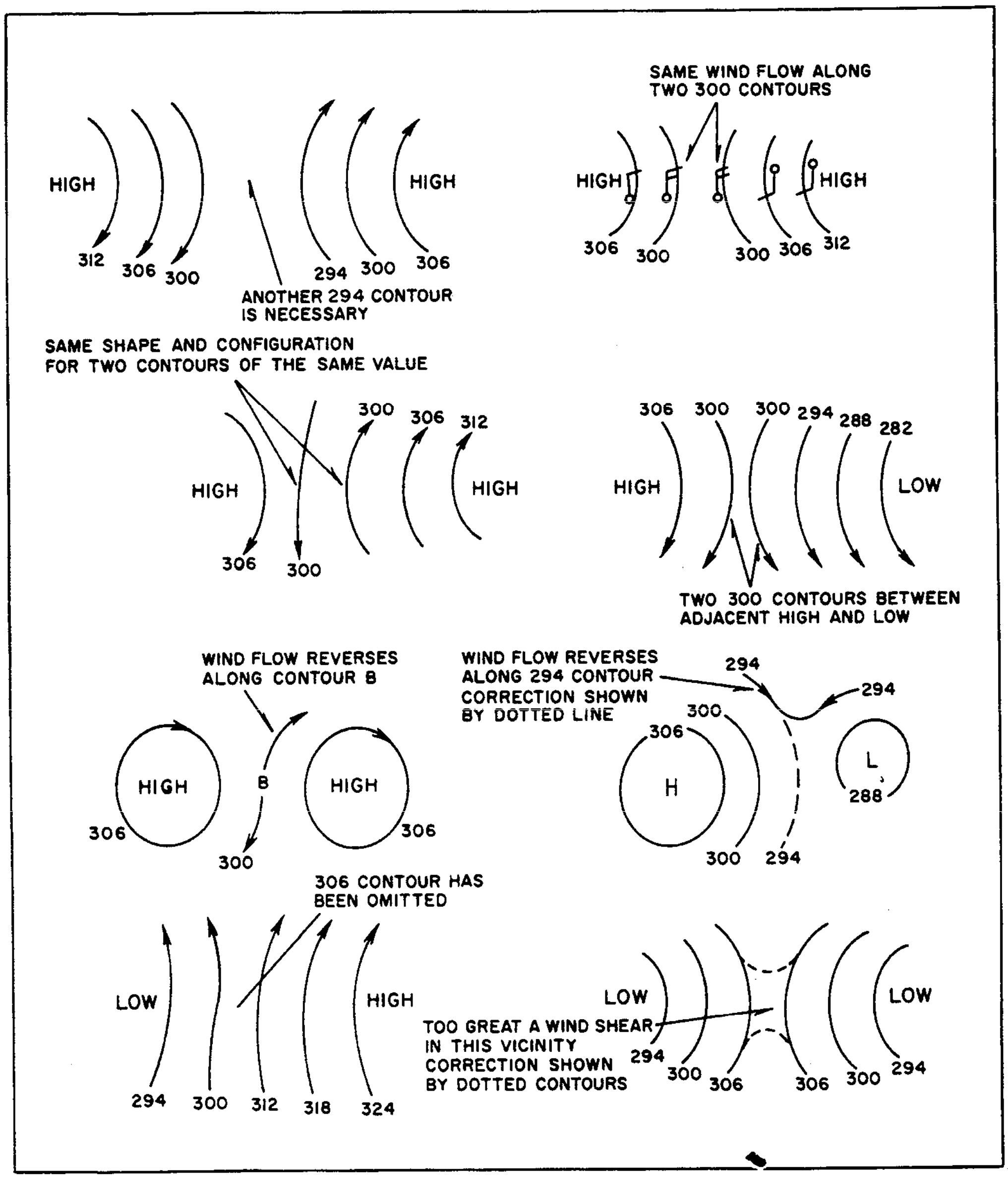

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Figure 8-1-5.—Common errors in contour analysis (700-mb chart).

200-mb chart, the ten-thousands digit is also dropped. Like isobars, contours are labeled wherever they end. If a contour extends from one side of a chart to another, it will be labeled at both ends. Closed contours are broken at a convenient spot to permit entry of the label. The heights should be labeled neatly and the labeling be of uniform size.

Upper-level contours form closed centers and wavelike patterns. The upper-level trough is an elongated area of low heights. Its axis is denoted by a trough line, which is sketched in black.

Upper-level ridges are elongated areas of high heights. A black zigzagging line is used to denote the axis of a ridge. Additional information on troughs and ridges is discussed later in this chapter.

Evaluating the Location and Slope of Pressure Systems and Fronts

The mechanics of constant-pressure analysis require that all levels interrelate; vertical consistency between levels is a must. The procedure used to determine vertical consistency is one of overlaying charts. Upper-level analyses are superimposed on lower-level analyses to assure a consistent slope to the various features with height. This permits visual determination of how systems stack. In this way, violations of internal consistency are prevented. For example, by overlaying a 1200Z 850-mb chart over the surface chart of the same day and time, you'll see that some pressure systems stack vertically, while others have a distinct slope. Lows normally slope upward toward colder air (usually westward and poleward), and highs toward warmer air (usually westward and equatorward). Fronts slope upward toward colder air. These spatial relations are an absolute requirement for a proper three-dimensional representation of pressure systems and fronts.

Although computer-generated surface and constant-pressure charts do not depict fronts, fronts do intersect upper levels. A fast-moving cold front with a slope of 1:40 intersects the 850-mb level approximately 35 miles to the rear of the surface position. A slow-moving cold front (1:100) is found no more than 90 miles behind the surface position. These same cold fronts intersect the 700-mb level 70 and 175 miles respectively to the rear of the surface position. Warm fronts have a more gradual slope (1:150) and may not intersect the 700-mb level. At the 850-mb level, expect to find them no less than 135 miles ahead of their surface positions. If warm fronts do intersect the 700-mb level, do not look for them within 250 miles of their surface position.

A good policy to follow in locating fronts is to first locate them aloft, then use the surface data to determine their exact positions. When upper-air information is sufficiently complete, only fronts meeting the following two criteria are ordinarily significant enough to be carried: Fronts that are supported by the temperature and wind patterns at the 850-mb level (700-mb level over mountainous regions); and fronts that can be identified in upper-air soundings. We'll discuss frontal analysis and temperature (isotherm) pattern relationships later in this unit.

Finalizing the Contour Analysis

After vertical consistency between levels is checked and any necessary adjustments made, the contours can be darkened in. This is usually done using felt-tip markers. High and low height centers are denoted by the letters *H* and *L*, respectively.

ISOTHERM ANALYSIS

The most common upper-air pattern consists of alternating troughs and ridges. They form a series of waves which encircle each hemisphere. Occasionally, closed lows exist in the troughs and closed highs in the ridges (fig. 8-1-6). The lows are normally poleward and highs equatorward of the basic wind flow. All of these features reflect the vertical extent of features at lower levels, sea level in particular. Note the axes of the systems shown in figure 8-1-7. They are not vertical. As stated earlier, lows or troughs slope upward toward colder air, and highs and ridges slope toward warmer air. To determine the location of the colder and warmer air aloft, we do an isotherm (temperature) analysis for each level.

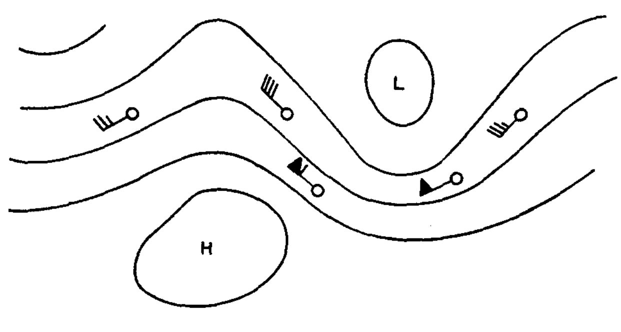

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Figure 8-1-6.—Common upper-air patterns.

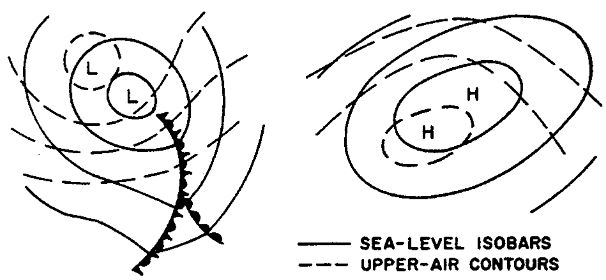

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Figure 8-1-7.—Slope of surface pressure systems with height.

Isotherm Patterns

Isotherm patterns at lower levels usually consist of tongues or pockets of warm and cold air that move across the map with reasonable continuity from day to day. There is a correlation between isotherm patterns and contour (isoheight) patterns.

Dynamically, warm tongues in the troposphere are related to pressure ridges. This relationship is due to the sinking and adiabatic warming of air, created by horizontally converging air at a higher level in the atmosphere. At the same time the air in the troposphere is sinking and warming, the convergence causes the air in the lower stratosphere to rise and cool. Intensifying upper-level ridges or highs are characterized by warming in the troposphere and cooling in the stratosphere. If a ridge builds (intensifies) and attains sufficient amplitude, the northern portion may cut off, forming a separate upper-level high.

Cold tongues are related to pressure troughs in the upper troposphere. This relationship is due to upper-level divergence and the resulting lifting and adiabatic cooling of air in the lower and middle troposphere. An additional consequence of this divergence is the sinking and adiabatic warming of air in the lower stratosphere. Deepening lows tend to be associated with intensifying cold tongues in the troposphere and warm tongues in the stratosphere.

Drawing Isotherms

Isotherms (lines of equal temperature) are usually drawn at 5°C intervals. They are drawn as solid red lines. They are sketched according to reported temperatures. Use their previous positions (past history) as a guide. Isotherms are labeled in the same manner as contours—at the end of each line and in a small break near the top of closed loops.

In regions of sparse data an estimate of the isotherm pattern becomes necessary. Integrating past history (heights and temperatures) with the latest surface-pressure pattern aids greatly in determining the pattern aloft. As we discussed earlier, temperatures at selected points are frequently used in extrapolating the heights of various upper levels. Because wind is the advecting mechanism of warm and cold air, you should be aware of its effect and how it impacts the thermal pattern.

WIND-TEMPERATURE RELATIONSHIP.— Wind determines the pattern isotherms take and their speed of movement. Their speed is somewhat less than the wind speed. The factors working against the advection of isotherms with the speed of the wind are as follows:

1. The addition or subtraction of heat.
2. Adiabatic changes due to lifting or subsidence. As warm air is advected toward a cold region, it tends to be cooled, and cold air advected toward warm regions tends to be warmed. In the case of cold air spreading southward over warmer regions, it tends to sink (being more dense), which results in adiabatic warming. At the same time, instability results. Both processes slow the advance of the cold air. Cold air moving offshore over warm ocean currents can be slowed by as much as 50 percent. When warm air is advected over cold regions, it tends to be lifted. This results in adiabatic cooling. The lifting process is more important at intermediate levels aloft; however, both lifting and subsidence work against the advection of isotherms with the speed of the wind.

The temperature pattern should explain temperature changes at individual stations. Because wind is the advecting mechanism, when there is a change of wind with height, the isotherm pattern must be adjusted. The isotherms are oriented to conform to the shear vector between the isobaric layers where the wind changed. The shear vector is derived from the wind speed and direction at the top and bottom of the layer.

WIND SHEAR.— The vectoral rate of change of wind with respect to altitude is called vertical wind shear. It is determined by taking the vector difference between the reported wind at the top and bottom of the layer and dividing by their vertical separation. On the other hand, the vector difference between geostrophic winds at two levels is called the thermal wind. The thermal wind is not a wind that actually blows in the atmosphere; rather, it is the vector difference between winds at two levels. For example, if the 700-mb geostrophic wind is southwest at 50 knots and the 500-mb geostrophic wind is west at 80 knots, the vector difference, or thermal wind, is from the northwest at 58 knots. See figure 8-1-8.

In that example, the mean isotherms between 500 mb and 700 mb would be oriented in a northwest to southeast direction, with colder air to the north. In general, thermal winds parallel mean isotherms in a given layer, with colder air to the left as you look downstream (the direction toward which the wind is blowing). The spacing of the isotherms is made to conform to the magnitude of the shear vector. The greater the shear vector, the closer the spacing of the isotherms (tighter gradient) and the more rigorously the direction of the isotherms conforms to the direction of the shear vector.

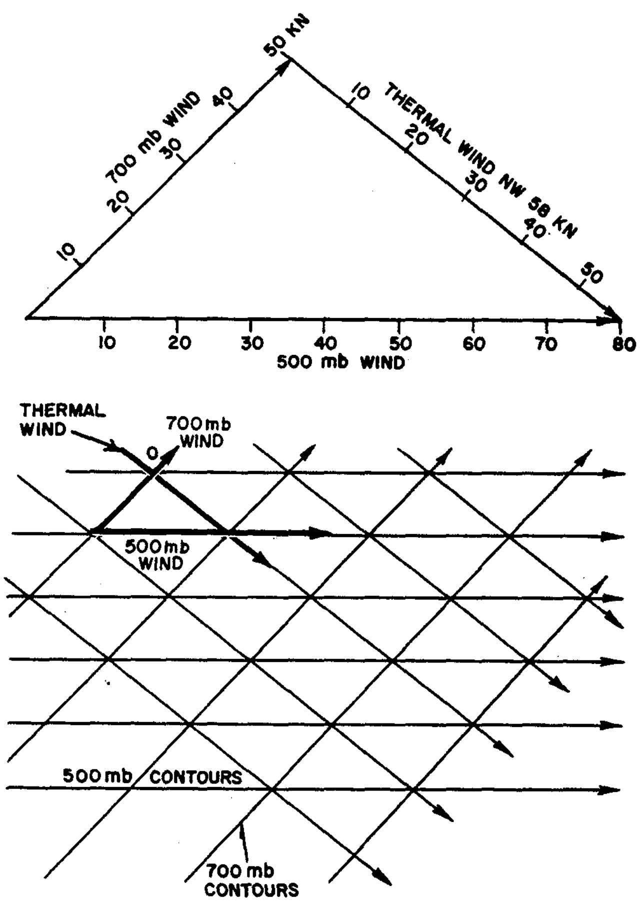

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Figure 8-1-8.—Thermal wind between two pressure surfaces.

Over the oceans, computed shear vectors are of considerable value in drawing isotherms. Speeds of less than 10 knots may not be significant, but when the shear vector exceeds this value, it is of particular use in the analysis.

In the free atmosphere the vertical wind shear is largely controlled by the temperature field. In areas where no temperature data is available, an isotherm analysis can still be carried out provided that the variation of wind with height is known. The basic relationships for the Northern Hemisphere are summarized as follows:

1. If wind speed increases with increasing height but does not change direction, contours and isotherms are parallel, with cold air to the left facing downstream.
2. If wind speed decreases with increasing height but does not change direction, contours and isotherms are parallel, with cold air to the right facing downstream.
3. If there is no change in wind speed or direction with height, the air temperature is uniform throughout the layer.
4. If the wind veers with increasing height, the isotherms cross the contours in such a way that advection of warmer air takes place.
5. If the wind backs with increasing height, the isotherms cross the contours in such a way that advection of colder air takes place.

FRONTAL ANALYSIS

The thermal gradient is by far the most important guide in locating fronts. The 850- and 700-mb isotherm analyses are used as aids in determining the location, strength, and vertical extent of fronts. Isotherms generally parallel fronts, with the tightest packing (gradient) in the cold air (fig. 8-1-9). On upper-level charts that do not show fronts, isotherms are continuous lines. On those showing fronts, the isotherms may be discontinuous.

The thermal gradient and pattern in relation to fronts give an indication of a front's strength. The weaker the gradient (in the cold air), the weaker the front and the greater the probability that the front has a shallow slope. The tighter the gradient, the stronger the front.

Isotherms may parallel cold fronts for long distances; however, somewhere to the north the isotherms may cross these fronts. Any appreciable crossing of these fronts by the isotherms indicates the front is occluded. Secondary cold fronts may also show an isotherm pattern that is more or less

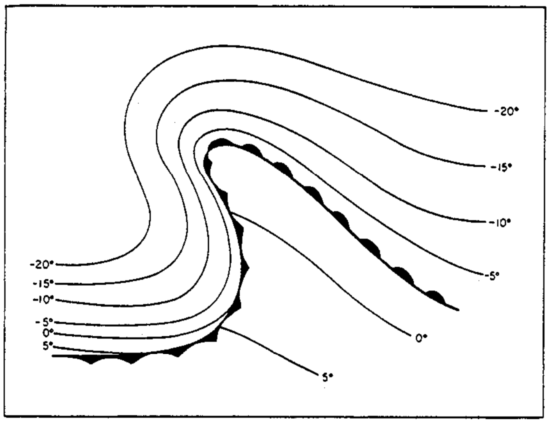

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Figure 8-1-9.—Isotherm packing at a warm and cold front aloft.

perpendicular to them. This indicates little air-mass contrast and a weaker front.

ISOTHERM-THICKNESS RELATIONSHIP

The isotherms of a constant-pressure level located near the middle of an isobaric layer are assumed to be representatiave of the mean isotherms (thickness lines) of the layer. For example, let's use the isobaric layer between 1000 mb and 500 mb. The constant-pressure level closest to the middle of the 1000-500-mb layer is 700 mb. The 700-mb isotherms are assumed to be representative of the mean temperature and thickness of the 1000-500-mb layer. Put another way, the 700-mb isotherm pattern is similar to the 1000–500-mb thickness pattern. The isotherm-thickness relationship applies to other layers. For example, the 850-mb isotherms are representative of the mean isotherms of the 1000-700-mb layer.

ISOTACH ANALYSIS

An isotach analysis provides a visual representation of the wind field at a given level. Isotachs are lines connecting points of equal wind speeds. They are drawn at 20-knot intervals. They form elongated ellipses that localize areas of highest wind speeds (fig. 8-1-10). These areas are known as centers of wind speed maxima in the overall isotach pattern. These centers move from west to east at speeds less than that of the winds themselves but greater than that of the waves in the long wave pattern. Their speed approximates that of the short waves. They move around the long wave troughs and ridges as these features move eastward. Occasionally, centers appear nearly stationary in the base of a trough or crest of a ridge, but they have

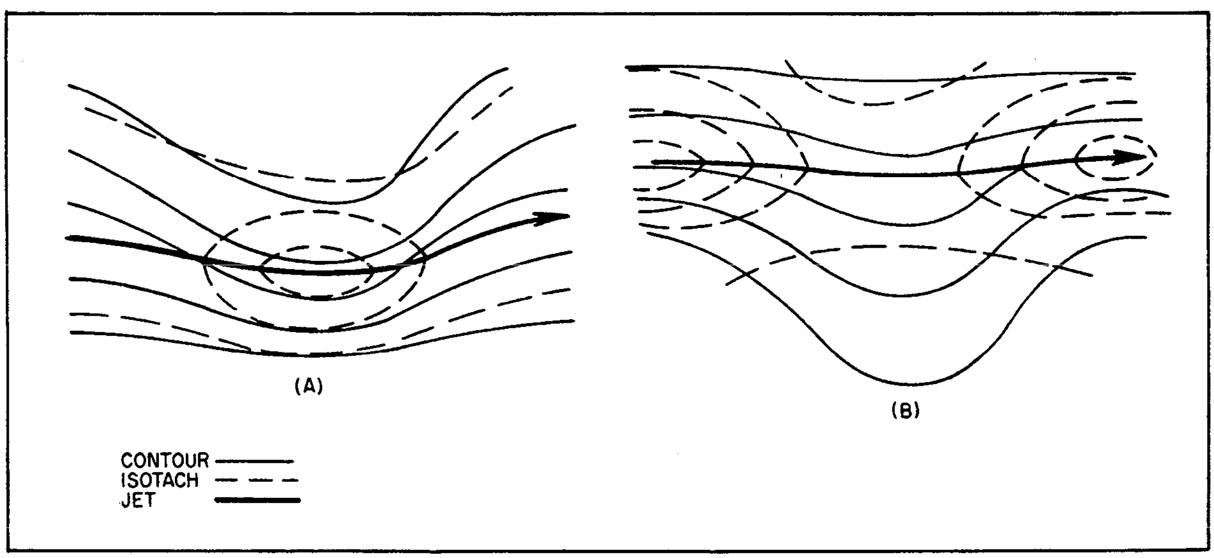

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Figure 8-1-10.—Common contour and isotach patterns. (A) Speed maximum at long-wave trough line; (B) Speed maximum at ridge line.

never been known to retrogress upstream. Isotach maxima on any constant-pressure chart represent the intersection of that level with a jet stream (jet).

Whatever level you analyze, begin the analysis in an area of dense reports. Sketch in the 60-knot isotach. Isotachs are drawn as a series of dashes. Begin at a point representing 60 knots and follow it downwind, interpolating between reported wind speeds. The isotachs will parallel, more or less, the height contours. However, at low speeds, isotachs cross contours at large angles, and where isoheights converge (upstream from speed maxima), isotachs tend to cross from high to low heights. The reverse takes place downstream, where isoheights diverge. Sketch in the remaining isotachs at 20-knot intervals and identify the centers of speed maxima. The centers are labeled with a "J" followed by the estimated maximum speed; e.g., J120. Speed minima (centers of minimum wind speeds) are labeled with an "S".

Computing wind speeds in sparse-data areas is accomplished using gradient or geostrophic wind scales. For reasons mentioned previously we use the geostrophic wind scale, and only in the case of cyclonically curved isoheights would you need to correct geostrophic wind to gradient wind.

JET STREAM ANALYSIS

A jet is defined as a narrow stream of relatively strong winds. On upper-level charts two such jets are often depicted. Polar-front jets are the most predominant and are associated with polar fronts of middle and subpolar latitudes. Like polar fronts, these jets can vary greatly in position and strength from day to day. The other jet that is often distinguishable is the subtropical jet. It is located at the poleward sides of the tropics, between 20 and 30 degrees latitude. To be classified as a jet, the wind speeds must equal or exceed 50 knots. However, an observation of high wind speed does not by itself warrant use of the term jet stream. As the word stream implies, the core must possess considerable length. The WMO defines it as normally thousands of kilometers in length, hundreds of kilometers in width, and some kilometers in depth.

The principal jet axis is marked by a heavy red line, with arrowheads indicating the flow direction. Secondary jets and jet fingers are depicted using dashed red lines with arrowheads. The jet axis, like isotachs, follows height contours. At 500 mb, it is located within 60 meters (plus or minus) of the 5,610-meter contour. At 300 mb, look for it between the 8,960- and 9,240-meter contours.

The polar-front jet lies vertically above the maximum temperature gradient of the middle troposphere. Because the 500-mb chart is representative of the mid troposphere, locate the maximum temperature gradient (isotherm packing) on this chart. The jet is normally found above and just south of this thermal band. This isotherm concentration is most often associated with a polar front, and the jet axis at 500 mb seems to coincide with the – 17°C isotherm on the warm side of the maximum temperature gradient.

The jet fluctuates vertically (up and down) and latitudinally (north and south). It usually lies between the 300- and 200-mb levels. It is nearer the 200-mb level at lower latitudes and 300 mb at higher latitudes. It also fluctuates seasonally. It is at a lower altitude in winter (300 mb) than in summer (200 mb).

The 300- and 200-mb charts give the best representation of the jet, but it is often well defined down at the 500-mb level. This level is very important in analyzing the jet, because it normally provides much better coverage (reports).

Upper-level charts allow us to locate jet streams and see their width and length. Another aid in locating their position is satellite pictures. The satellites play a key role over the oceans where upper level reports are very sparse. The combination of upper-level-wind observations and satellite pictures has given us a much better understanding of the various jet streams that exist within the atmosphere. There will be more on jet streams later in this chapter.

In summary, the isotach and jet stream analyses go hand in hand. Isotach patterns from 500 to 200 mb outline centers of wind speed maxima and their associated jet streams. The jet streams are rivers of fast-flowing air (50 knots or more) and extend for thousands of miles around Earth, are hundreds of miles wide, and have a variable depth of only a few miles. The jet core is usually located between 300 and 200 mb, but the entire stream fluctuates vertically and horizontally. A combination of upper-level-wind reports and satellite pictures provides us with a means of locating the jets occurring within our atmosphere.

TROPOPAUSE ANALYSIS

Unless you operate as a member of an upper air team taking raobs and rawinsondes, you will most likely never do a tropopause analysis. The main things to remember are the relationship between the tropopause and the troposphere, and the criteria an analyst uses in determining where the troposphere ends and the tropopause begins.

The tropopause is a transition boundary between the troposphere and the stratosphere. The analysis of this boundary layer is carried out on adiabatic charts, using temperature data obtained from atmospheric soundings (raobs and rawinsondes). The tropopause is characterized by an abrupt change in the temperature lapse rate. The change is from a lapse rate marked by decreasing temperature through the troposphere to one where the temperature decreases much more slowly, becomes isothermal (constant), or even shows a slight increase with height through the tropopause.

There are at least three distinct tropopauses, which form leaflike or overlapping structures within Earth's atmosphere. The three most generally accepted are the subtropical tropopause, found at 25° latitude near 18,290 meters (approximately 100 mb); the mid-latitude tropopause, at 35° to 40° latitude near 12,190 meters (approximately 200 mb); and the subarctic tropopause, near 9,145 meters (300 mb). In general, each tropopause is found at greater heights in summer than in winter. Because the thickness of the troposphere increases from the poles to the equator, each tropopause slopes upward toward the equator. For a short distance, the subtropical tropopause tends to overlap the mid-latitude tropopause, and the mid-latitude tropopause tends to overlap the subarctic tropopause. Figure 8-1-11 illustrates the overlapping of the mid-latitude and subtropical tropopauses.

Each tropopause is characterized not only by height and pressure but also by potential temperature. In winter, the potential temperature of the subtropical tropopause is, ±10, 390K; the mid-latitude tropopause, 350K; and the subarctic, 310K. Potential temperatures are used to locate tropopauses from atmospheric soundings with many inversions or irregular lapse rates with no inversions.

The variation in temperature structure through the tropopause led to many different ideas on how to define and analyze it. The present WMO definition does not state what a tropopause is. Instead, the WMO defines an objective technique (selection criteria) for locating tropopauses from

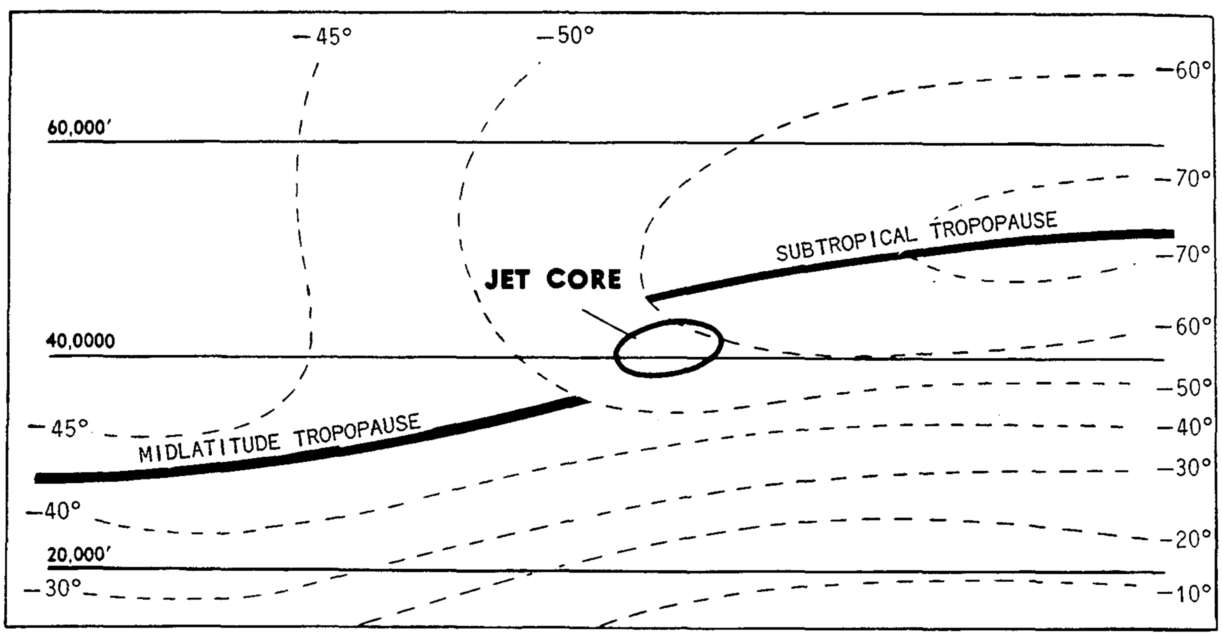

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Figure 8-1-11.—Overlapping of subtropical and mid-latitude tropopauses.

a plotted atmospheric sounding. The significance of WMO's definition is the standardization the definition has brought about. It permits the same height to be consistently selected by all technicians from a given sounding, and other atmospheric phenomena can be empirically (through experience and observation) related to it.

Within the WMO definition, provision is made for identifying two or more tropopauses on a sounding, which is necessary because of the regions where overlapping exists. The definition is in two parts.

1. The "first tropopause" is defined as the lowest height at which the lapse rate decreases to 2°C per kilometer or less, provided also that the average lapse rate between this height and all higher altitudes within 2 kilometers does not exceed 2°C per kilometer.
2. If, above the first tropopause, the average lapse rate between any height and all higher altitudes within a 1-kilometer intend exceeds 3°C per kilometer, then another tropopause is defined by the same criteria as under 1 above. This second tropopause may be within or above the 1-kilometer layer.

There are also two qualifying remarks attached to the selection criteria. They are as follows:

1. A height below the 500-mb level is not designated as a tropopause unless the sounding reaches the 200-mb level and the height is the only height satisfying the above definitions.
2. When the second or higher tropopauses are being determined, the 1-kilometer interval with an average lapse rate of 3°C per kilometer can occur at any height above the conventional tropopause and not only at a height more than 2 kilometers above the first tropopause.

SUPPLEMENTARY UPPER-AIR ANALYSIS

The basic upper-air analysis is the constant-pressure analysis. In conjunction with this basic analysis it is sometimes necessary, and at all times beneficial, to conduct concurrent supplementary types of analyses of upper-air properties in order that the fullest use be made of upper-air information to lead to the end product, the forecast. Most of these charts are constructed from either reported or derived data from upper-wind and upper-air reports. It is not feasible to list or explain all the types of upper-air charts currently being produced by the National Meteorological Center. Only the space differential (thickness) analysis, advection, and time differential charts are covered here.

Space Differential (Thickness) Analysis

For a truly three-dimensional analysis of the free atmosphere, it is necessary to analyze not only the individual levels within the atmosphere (850 mb, 700 mb, etc.) but also the various layers of it. The most commonly analyzed layers are the 1000-700-mb, 1000-500-mb, 700-500-mb and 500-200-mb layers. Layer, or differential, analysis ensures vertical consistency between the individual levels and agreement within the limits of the hydrostatic equation.

Know the heights of any two constant-pressure levels and you can determine the thickness (vertical distance) of the layer separating them. It's simply a matter of subtraction. Space differential charts are commonly referred to as thickness charts, since they represent the difference in height between two constant-pressure levels. Manually, they are constructed by graphically subtracting the heights of one analyzed constant-pressure chart from those of another.

PROCEDURE FOR ANALYSIS.— If the need should ever arise whereby you must do such an analysis, here's a recommended procedure:

1. Obtain the analyzed constant-pressure charts for the levels bounding the layer to be analyzed.
2. Place an acetate over one of these charts and trace the isoheights onto the acetate with a grease pencil. Be sure to label the contours.
3. Place the same acetate over the other chart and trace and label the isoheights of this level, using a different color grease pencil.
4. Graphically subtract the lower contours from the upper and connect the points of equal thickness. It may prove helpful if you perform the mathematical subtraction of heights at a few intersections to aid you in getting the analysis started. Thickness lines are drawn in dashed black.
5. Place the acetate under a clean chart and trace the thickness lines onto the chart with a black felt-tip pen. Label the thickness lines.

There are several rules that you must follow in constructing the thickness pattern:

1. Thickness lines cannot touch or cross.

2. Thickness lines cross isoheight contours at the intersection points of the two levels only.

3. Thickness lines must always pass from lower to higher contours or vice versa at both levels.

4. Between any two consecutive thickness lines an isoheight of either pressure surface must exist.

5. Between any two consecutive isoheights a thickness line or an isoheight of the other pressure surface must exist.

THICKNESS PATTERNS.— Figure 8-1-12 illustrates most of the important details of the 1000-500-mb thickness pattern in relation to fronts. Adherence to these features of the thickness model insures the proper slope of systems between 1000 and 500 mb, the proper relationship between surface fronts and polar jet, and surface frontal analyses that portray a meaningful picture of the three-dimensional temperature structure. The most important features of the pattern are as follows:

1. The concentration of thickness contours is on the cold side of frontal systems. The stronger the front, the greater the concentration.

2. The spacing of thickness contours in the cold air ahead of warm fronts is greater than in the cold air behind cold fronts.

3. The horizontal distance between the maximum thickness gradient and cold fronts is less than with warm fronts. The maximum gradient is usually located horizontally in the same position as the 500-mb jet stream.

4. Thickness contours are anticyclonically curved in advance of warm fronts and cyclonically curved behind cold fronts.

5. The location of the cold trough in the thickness contours lies to the rear of the surface low, halfway between the surface low and the next upstream ridge or high.

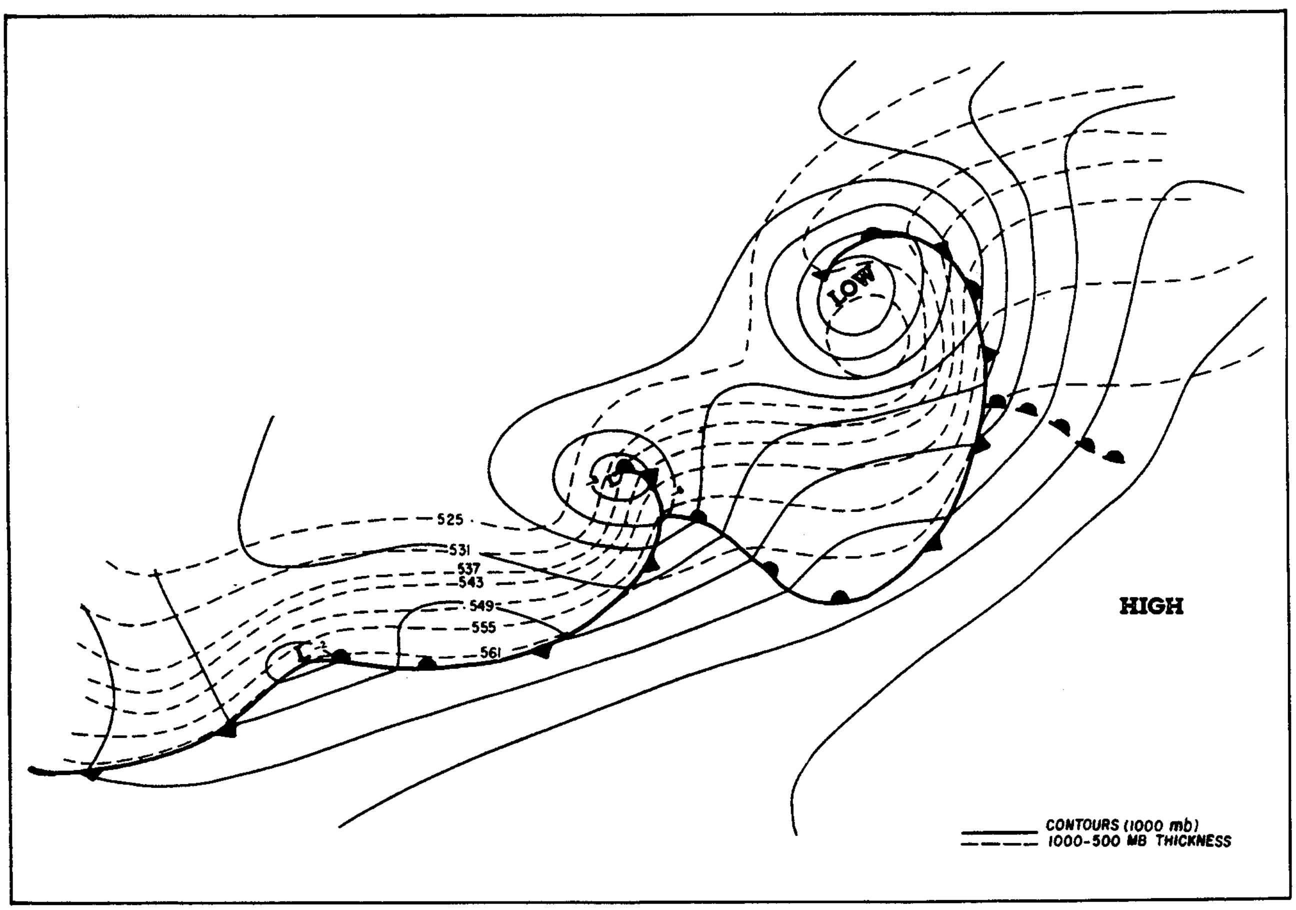

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Figure 8-1-12.—Illustration of the relation of thickness patterns to fronts.

Advection Analysis

In meteorology, advection refers to the horizontal transport of heat or other properties. Advection charts, like thickness charts, are constructed between any two desired constant-pressure levels. In fact, they are usually constructed on thickness charts. Cold or warm air advection is indicated at the same intersections used to compute thickness.

Advection arrows indicate the mean direction of flow in the layer. They are, for the most part, perpendicular to the thickness lines. Blue arrows indicate cold air advection and red arrows warm air advection. If the flow is such that it crosses thickness lines from higher height values to lower height values, the advection is warm; and from lower to higher values, it's cold. The thumb rule to follow when constructing advection arrows is that when the contour of the lower pressure level is to the left looking downstream, advection is cold; but when it is to the right, advection is warm. See figure 8-1-13.

A derivative of the thermal wind equation and advection arrows is the thermal wind rule. It simply states that in the Northern Hemisphere if the wind backs with height, cold advection is indicated; and if the wind veers with height, warm air advection is indicated. Figure 8-1-14 illustrates the thermal wind rule as it applies in the Northern Hemisphere.

Time Differential Analysis

Time differential charts show the amount and direction of change of a meteorological quantity over a given period of time. They are used to track upper-height centers. They are usually drawn at 24-hour intervals, vice every 12 hours, to minimize diurnal effects.

Construction is as follows: Take two charts, 24 hours apart and of the same level, and lay the most recent over the other. Over these lay a clean acetate or blank chart. Using a light table, algebraically subtract the contour values of the most recent chart from those of the earlier chart. This is done where the contours of the two charts intersect. Lines are then constructed connecting like height differences. Falling height lines are drawn in red, rising heights in blue, and lines of no height change are drawn in purple. These lines are labeled in decimeters. The centers of rising

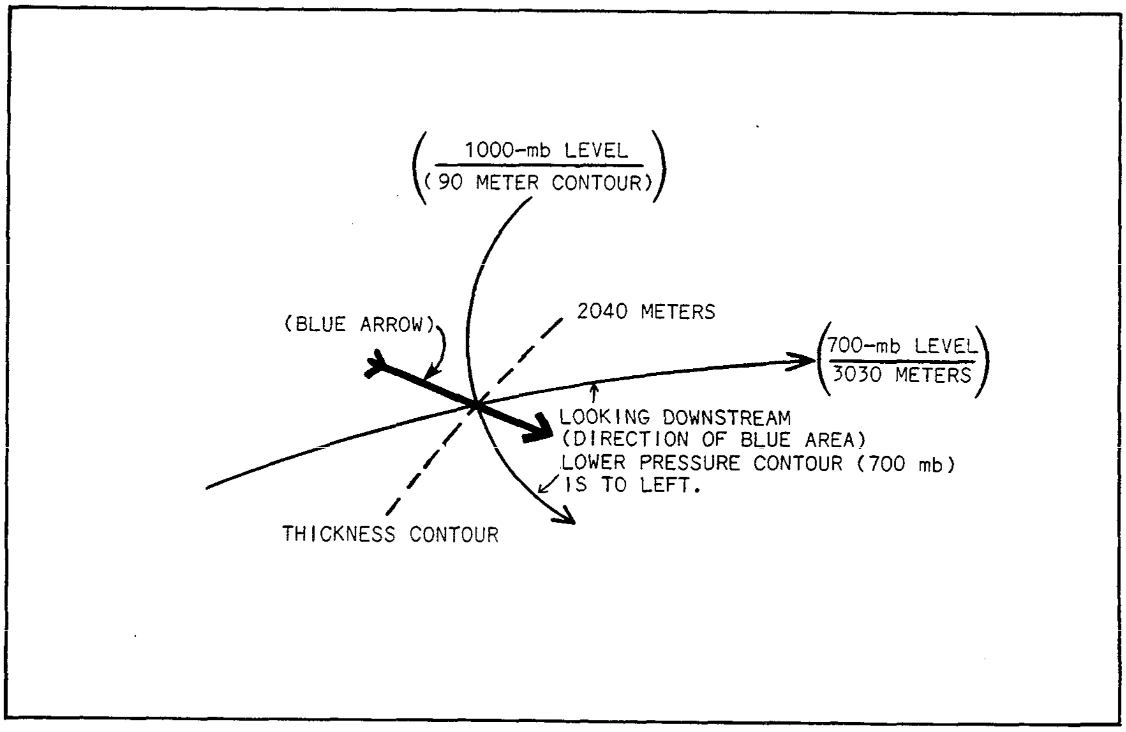

Figure 8-1-13.—Explanation of advection arrow construction.

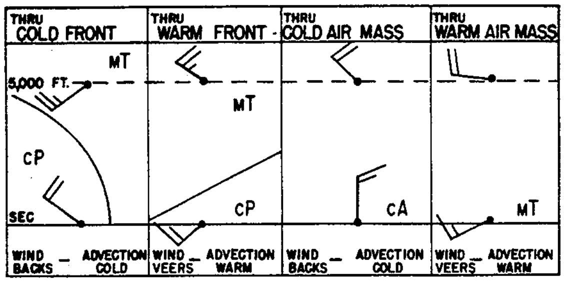

Figure 8-1-14.—The thermal wind rule (Northern Hemisphere).

and falling heights are then transposed onto a blank chart (if acetate was used). These centers are then tracked in the same manner as are the pressure centers on constant-pressure charts. See figure 8-1-15 for an illustration of time differential construction.

Summary of Upper-Air Analysis Rules

The following list is by no means all inclusive, but it does provide a guide for avoiding some typical and often repeated errors made by inexperienced analysts.

1. Use history. Always check the previous analyzed map before starting the current analysis.

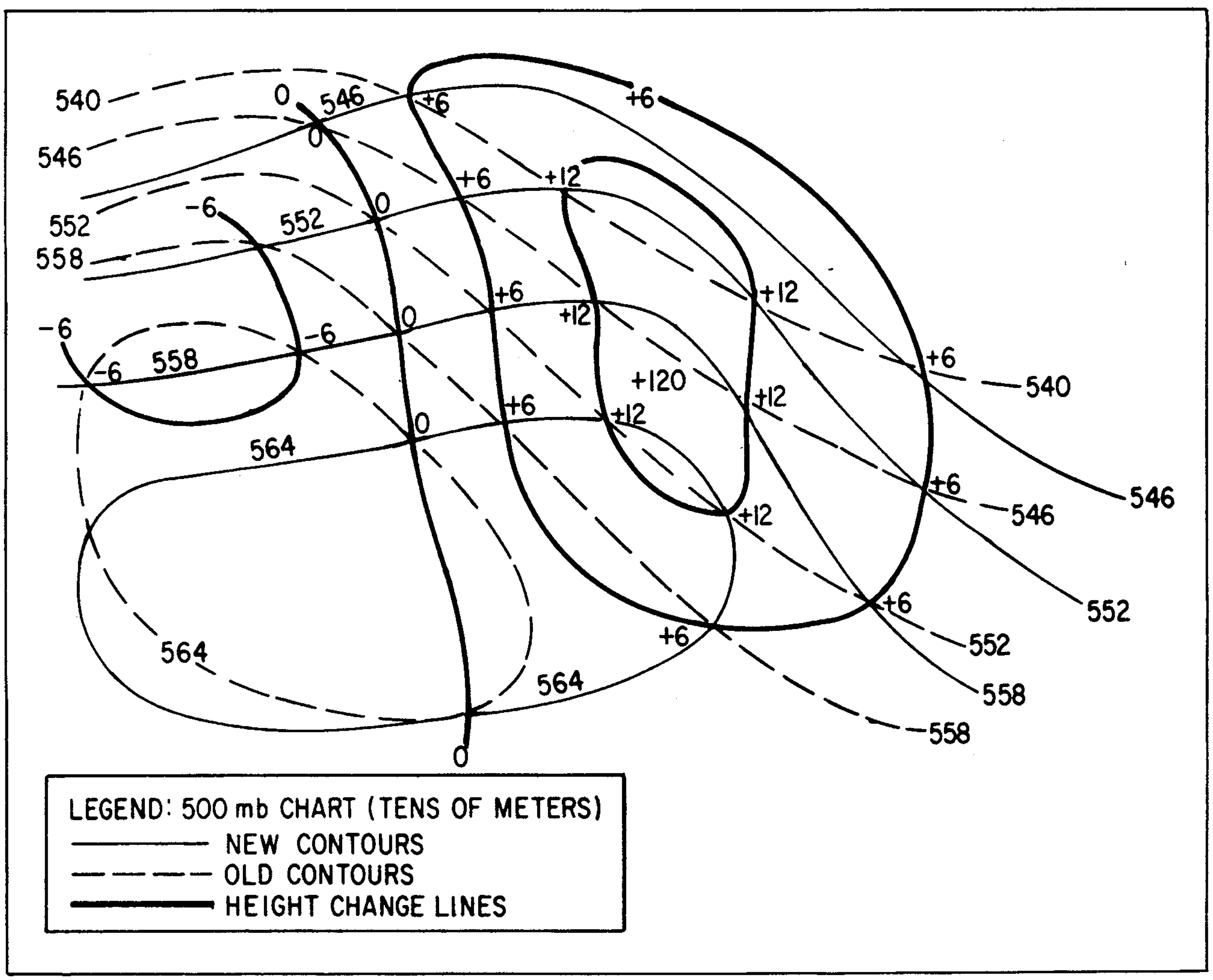

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Figure 8-1-15.—Time differential chart.

2. Isoheights (contours) are drawn parallel to winds where possible. This is not always possible, because of observation errors and nongeostrophic and/or nongradient winds. Always check the plotted wind shaft for the number indicating direction.

3. Contours are drawn following down the wind; therefore, wind direction cannot change discontinuously along a contour.

4. Do not overemphasize cyclonic curvature at troughs in the form of kinks. This implies a trough is a front.

5. Continue analysis into areas of no data. Use wind scales, history, and common sense to get central heights of highs and lows when no data is available.

6. Use geostrophic wind scales to determine correct spacing of contours. The spacing will be less than the observed wind indicates in areas of cyclonic curvature and greater than the observed wind where anticyclonic curvature is occurring.

7. The isotherm-contour pattern at 500 mb is normally one of cold lows and warm highs. A closed isotherm will be nearly coincident with height centers. Long-wave troughs are cold, ridges warm. The opposite is true for short waves.

8. Marked changes in the configuration of troughs should be doubted.

9. Insure vertical consistency.

a. Troughs and lows must slope toward coldest air.

b. Ridges and highs must slope toward warmest air.

c. Cold lows have little or no slope.

d. Warm lows at lower levels become short-wave (warm) troughs aloft.

e. Frontal waves at sea level become short waves aloft.

f. Occluded lows become cold or cutoff lows aloft.

10. Strive for a professional-looking product. Avoid jagged, ragged or nervous isolines.

UNIT 8—LESSON 2

USE OF CONSTANT-PRESSURE CHARTS

OVERVIEW

Identify the uses for the 1000-, 850-, 700-, 500-, 300-, 200-, 150-, 100-, 50-, and 25-mb constant-pressure charts.

OUTLINE

The 1000-mb chart

The 850-mb chart

The 700-mb chart

The 500-mb chart

The 300-mb chart

The 200-mb chart

The 150-, 100-, 50-, and 25-mb charts

CONSTANT-PRESSURE CHARTS

Upper-air charts are termed CONSTANT-PRESSURE CHARTS, because they depict conditions at levels (heights) within the atmosphere where the pressure is the same (constant).

Constant-pressure charts are produced for standard levels. These levels are the 1,000-, 850-, 700-, 500-, 400-, 300-, 250-, 200-, 150-, 100-,70-, 50-, 30-, 20-, 10-, 7-, 5-, 3-, 2-, and 1-mb levels. However, the most commonly produced charts are for the 1000-, 850-, 700-, 500-, 300-, 200-, and 100-mb levels.

Atmospheric soundings show that pressure changes most rapidly when the temperatures are cold and least rapidly when they are warm. Remember, pressure is a function of the weight of the atmosphere, and the atmosphere's weight is dependent on its density. Cold air is more dense than warm air; therefore cold air is heavier and exerts more pressure at a given altitude. Assuming that two columns of air (one cold, one warm) exert the same pressure at the surface, the column containing the warm air has to extend to a greater height. Figure 8-2-1 illustrates the pressure-height differences in cold and warm air. Also, note the vertical spacing (thickness) between the constant-pressure levels.

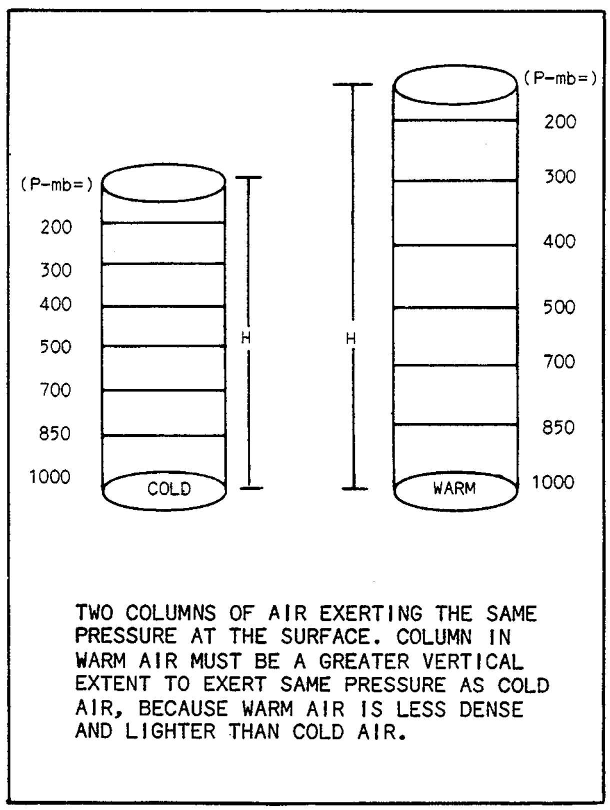

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Figure 8-2-1.—Pressure-height differences in cold and warm air.

As the temperatures in the atmosphere change, so do the heights of the constant-pressure levels. Based on the U.S. standard atmosphere, the approximate heights of the more common constant-pressure levels are as follows:

Pressure level (millibars)	Heights (meters)	(feet)
1000	110	370
850	1,460	4,780
700	3,010	9,880
500	5,570	18,280
400	7,180	23,560
300	9,160	30,050
200	11,790	38,660
150	13,620	44,680
100	16,210	53,170

Constant-pressure charts are primarily used as an aid in weather forecasting. When they are used in conjunction with surface synoptic charts, the following determinations may be made:

1. Movements of weather systems
2. Areas of cyclonic and anticyclonic windflow
3. Types of air masses
4. Location of moist and dry areas within the atmosphere
5. Formation, intensification, and dissipation of pressure systems
6. Actual slopes of fronts
7. Vertical extent of pressure systems
8. Location and strength of jet streams

Learning Objective: Identify the uses of constant-pressure charts commonly prepared and displayed in most weather offices.

The following is a brief summary of the principal uses of constant-pressure charts commonly prepared and displayed in most weather offices.

THE 1000-MB CHART

This chart indicates the height of the 1000-mb level above and below sea level. When below sea level, it is indicated by negative height values. For reasons previously discussed, 1000-mb height values are not very accurate in mountainous areas. Use them with caution.

A 1000-mb chart is constructed from a surface chart by assuming that 7 1/2 mb equals 60 meters for temperatures between 30°F and 70°F; that 8 1/2 to 9 mb equals 60 meters for temperatures below 30°F; and that 6 1/2 mb equals 60 meters for temperatures above 70°F. These ratios give valid approximations of 1000-mb heights; however, heights from radiosonde soundings should be used whenever available.

The principal use of 1000-mb chart is in constructing space differential (thickness) charts. The 1000-mb chart serves as the base level for the 1000-700- and 1000-500-mb charts.

THE 850-MB CHART

The principal uses of the 850-mb chart are to locate surface frontal positions, to determine the representativeness of surface winds and temperatures, to determine the depth of moisture patterns in winter, and to serve as the surface chart in mountainous and plateau areas where the mean elevation is around 5,000 feet.

Temperature and moisture analyses should be made in close conjunction with the surface chart whenever possible. A complete and careful isotherm analysis at this level in conjunction with the wind and height analysis will lead to the correct placement of fronts both at this level and the surface. A thumb rule to guide you in locating most fronts is to look for the 850-mb warm front roughly 2 1/2 to 3 degrees latitude ahead of the surface front and cold fronts 3/4 to 2 degrees latitude behind the surface front. An isotherm analysis also serves as a good indication of the 1000-700-mb thickness pattern.

The isoheight interval is the same as that used by the National Weather Service, 30-meters.

THE 700-MB CHART

The 700-mb chart is used mostly to determine the vertical extent and structure of fronts and pressure systems, or to play the role of the 850-mb chart over areas where the mean elevation is around 10,000 feet. It is also used to analyze moisture patterns in summer, when moist tongues extend to greater heights than in winter, because of convective activity. Other uses are in forecasting (steering currents for certain shallow pressure systems are determined at this level) and differential analysis.

Short waves are a predominant feature of this chart. Because of their influence on cloudiness, frontal intensity, precipitation areas, etc., these wave features are carefully studied and tracked. The isoheight interval is the same as that for the 850-mb chart-30 meters.

THE 500-MB CHART

The 500-mb chart is the most widely used of all upper-air charts. Primary features are the warm highs and cold lows, with their associated ridges and troughs. Long waves are identifiable at this level, but most short waves have lost their identity.

For various reasons, the 500-mb level comes closest to representing the mean state of the atmosphere at observation time. Since this level approximately divides the atmosphere with respect to mass, the 500-mb chart is often used, in conjunction with the 1000-mb chart, to provide a layer analysis of the lower half of the atmosphere.

This chart also provides winds at a common flight level for piston-engine aircraft, gives a fair approximation of the horizontal position of the jet stream when no 200- or 300-mb chart is available, and provides an important base upon which to construct higher-level analyses. The importance of this last point is based on the rapid decrease in available data above 500 mb. The 500-mb chart is also used extensively in forecasting the movement and development of sea level systems and fronts. Contour spacing is 60 meters.

THE 300-MB CHART

The primary features of the 300-mb chart are the permanent and semipermanent highs and lows, certain dynamic lows, long waves, the polar jet stream in winter, and the tropopauses, especially the arctic and mid-latitude tropopauses in winter.

Its primary uses are in forecasting; determining the characteristics of long waves; analyzing and forecasting jet streams; analyzing the tropopause in winter; determining vorticity distribution; and in the case of tropical lows that do not show a closed circulation at this level, steering currents. It is also an indispensable tool in planning jet aircraft operations.

The contour interval is normally 120 meters, but a 60-meter interval maybe used in areas where a finer degree of delineation is required.

THE 200-MB CHART

The operational use and contour interval of the 200-mb chart are the same as those of the 300-mb chart. In fact, the 200-mb chart is used as an adjunct to the 300-mb analysis. In summer, it plays the same role, with respect to the jet stream, that the 300-mb chart does in winter. In winter, its principal use is in estimating changes in the temperature advection pattern in the stratosphere.

THE 150-, 100-, 50-, AND 25-MB CHARTS

The 150-, 100-, 50-, and 25-mb charts are normally prepared at major centers and are used primarily for research purposes. Data at these levels is so scanty that their construction is almost solely based on extrapolation of data from lower levels. As the operational ceilings of jet aircraft increase, it can be anticipated that practical uses of the 150- and 100-mb charts will also increase.

UNIT 8—LESSON 3

CIRCULATION PATTERNS ON UPPER-AIR CHARTS

OVERVIEW

Identify atmospheric circulation patterns and define special circulation features.

OUTLINE

Long and short waves

Upper-level highs and lows

Blocks

Zonal and meridional flow

Jet streams

CIRCULATION PATTERNS ON UPPER-AIR CHARTS

The patterns on constant-pressure charts are much the same as those patterns delineated by isobars on surface charts. However, the pattern becomes smoother and less complex on upper-level charts. The 850-mb chart, for example, may show many closed centers, but by the time the 200-mb level is reached, there are few, if any, closed centers. Instead, the contours present a wavelike pattern. These waves, like their ocean counterparts, have definite properties by which they can be identified.

Learning Objective: Identify the circulation patterns of long waves, short waves, upper highs, and upper lows on upper-air charts.

LONG AND SHORT WAVES

Waves are classified according to their length, amplitude, and speed. Wavelength is the measured distance (in degrees longitude) between successive waves. The measurement is usually taken from trough to trough, ridge to ridge, or from any point on one wave to the same corresponding point on the next wave. The amplitude is one-half of the wave's total range, which is measured in degrees latitude from the peak of the ridge to the base of the trough. Figure 8-3-1 illustrates the measurement of wavelength and amplitude on a long wave. Also, note the short wave on the long wave. The speed of waves is usually governed by their length. The longer the waves, the slower they move, and vice versa.

Long Waves

A significant feature of the westerlies in both hemispheres, long waves vary in length from 50° to 120° longitude, have large amplitudes, and are

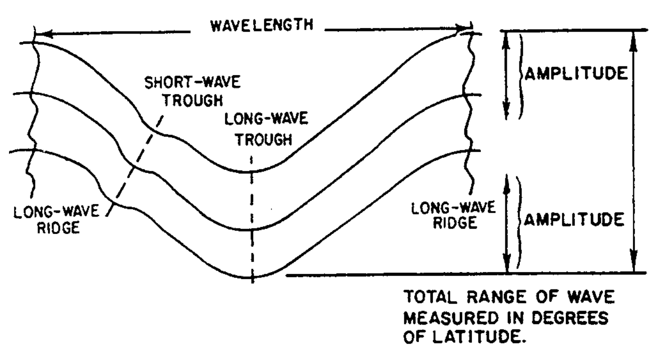

Figure 8-3-1.—Illustration of a long and short wave and the measurement of length and amplitude of a long wave.

slow moving. In the overall hemispheric pattern there are normally four or five long waves in existence at any one time. However, there are times when there are as many as seven or as few as three. The pattern is a persistent feature, and waves do not appear or disappear rapidly. A change in the number of waves in the pattern is significant. The fewer the number, the more progressive are the weather patterns at the surface. The greater the number, the more stagnant the weather patterns. It is during these periods that prolonged good or bad weather affects a region. The number of waves and pattern changes are often discussed at briefings for meteorologists. New long waves form from short waves or a changing synoptic situation, and their development is associated with the development of new intense circulations at lower levels.

Because some short waves have large amplitudes, it is often difficult to distinguish them from long waves. Also, it is virtually impossible to identify wave types (long or short) on a single chart. A series of charts over a 3- to 5-day period is best for this. This is usually enough time for most short waves to move through the slower long-wave pattern, thereby distinguishing between the two types. The long waves have a normal movement at 40°N of about 2° longitude per day in the spring to less than 1° per day during the fall. They can also become stationary or even retrogress.

Because a long wave's amplitude increases with height within the troposphere (greater at 300 mb than at lower levels) the long-wave pattern is best identified at the 300-mb level. Here, the wave contours are approaching their maximum amplitude, and the overall pattern is smooth (no short wave distortion). The increase in amplitude with height also distinguishes long waves from short waves. Short waves often disappear with height and may not be detectable above 500 mb. This is attributable to the temperature patterns associated with the two types of waves. With long waves, the troughs are cold and the ridges warm; the opposite holds true for short waves.

Synoptically, long waves are related to a number of weather occurrences. Figure 8-3-2 illustrates many of the relationships. Simply stated, the weather between the long-wave trough and the downstream ridge (trough sector) is more apt to be bad than the weather between the trough and the upstream ridge (ridge sector).

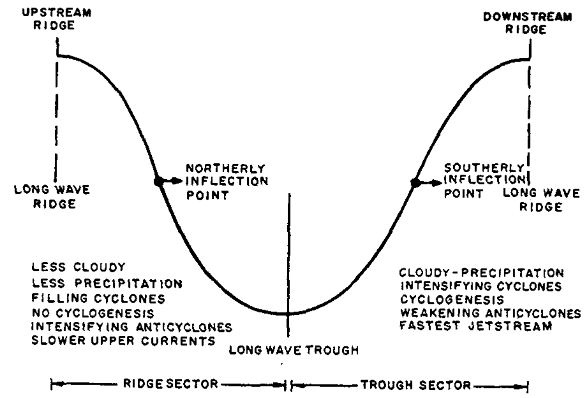

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Figure 8-3-2.—Long waves and related weather.

Short Waves

Superimposed on the long-wave contours of a given upper-air chart, say 500-mb, are numerous short waves. Ten or more short waves are present in the hemisphere most of the time. They have shorter wavelengths and smaller amplitudes and move faster than long waves. They move in the same direction as the current in which they are embedded. Their eastward motion is very near that of the 700-mb flow. Their normal movement is on the order of 8° longitude per day in summer and 12° per day in winter. Short waves are progressive and never retrograde. Their troughs are warm and their ridges cold; therefore, they do not extend to great heights and are most predominant in the lower half of the troposphere (500 mb and below).

Short waves have a great effect on long waves. They dampen (flatten) long-wave ridges as they move across them, and at the same time, the short wave is weakened. As a short wave approaches a long-wave trough, the short wave strengthens and the long-wave trough intensifies. This latter occurrence often results in the formation of a surface low-pressure system (cyclogenesis).

The location of short waves coincides with the small closed height fall centers (troughs) and height rise centers (ridges) of a 700- or 500-mb time-differential chart.

Wave Movement

We've discussed the relationship between waves and their associated temperature patterns. That is, long waves have cold troughs and warm ridges, while short waves have the opposite. The location and strength of warm and cold air in relation to troughs and ridges give an indication of their present and future movement. Locating and determining the strength of waves is done through a comparison of the isotherm and contour patterns on constant-pressure charts.

The isotherms are considered either in phase or out of phase with pressure troughs and ridges. In phase, the thermal troughs and ridges coincide with the pressure troughs and ridges. Figure 8-3-3, cases A through C illustrate the in-phase relationships. The only difference in the three in-phase patterns is the amplitude of the isotherms as compared to that of the contours.

When the isotherms and contours are in phase and parallel (have the same amplitude), the wave stagnates, because there is no temperature advection taking place across the wave.

When the isotherm amplitude is less than the amplitude of the contours (case B), colder air is advected into the west side of the trough and warm air into the eastern side. Where the temperatures are falling (in this case, the western side), the pressure should fall. Since systems move from high to low pressure, we would expect such a wave to move west (retrograde).

When the isotherms have a greater amplitude than the contours have (case C), warm air is advected into the western side of the trough and colder air into the eastern side. Therefore, pressures fall in advance of the trough and rise behind it. When this occurs, the trough moves slowly eastward toward the falling pressures (is slowly progressive).

The last two cases deal with out-of-phase isotherm-contour relationships and are associated

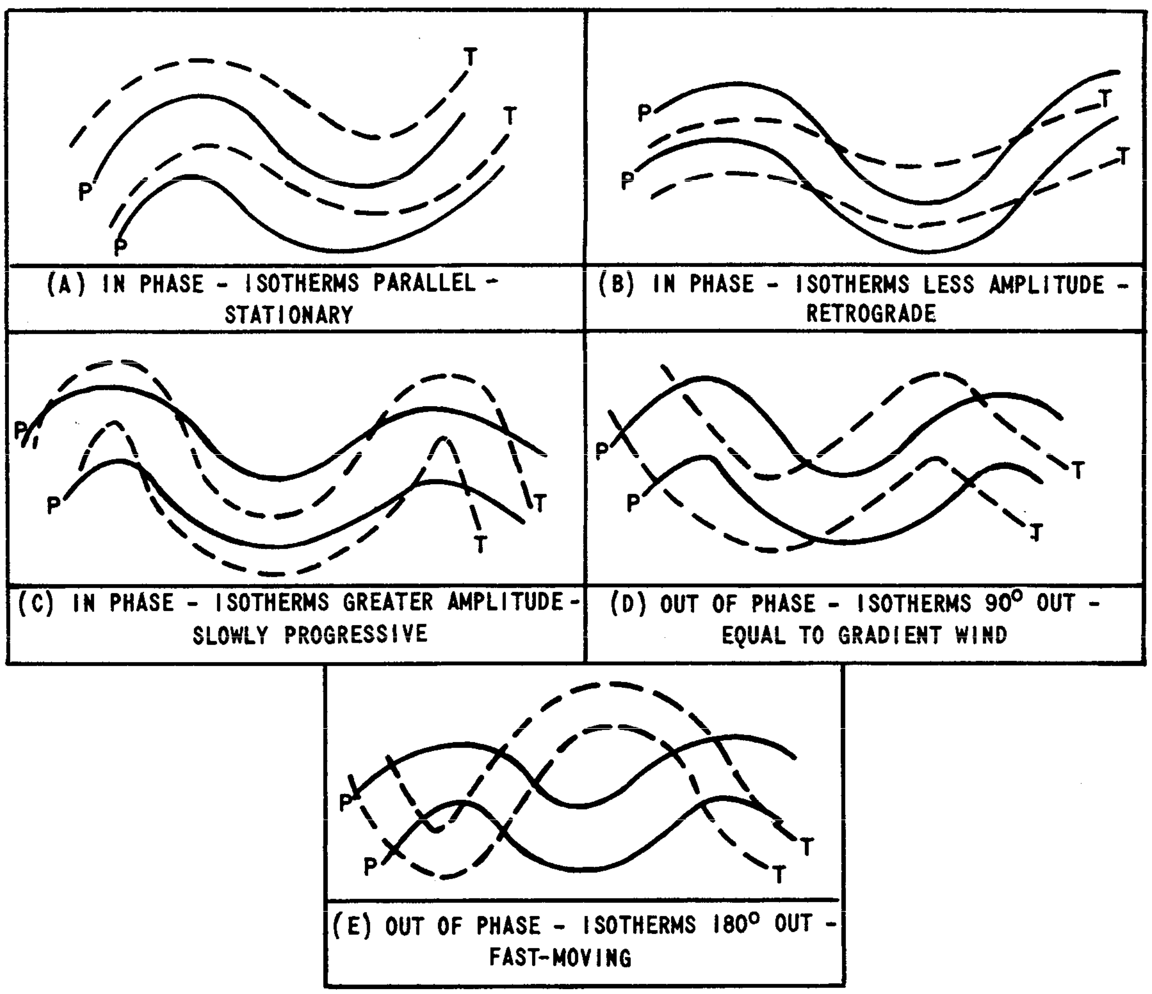

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Figure 8-3-3.—Isotherm-contour patterns.

with short waves. See figure 8-3-3, cases D and E. These isotherms are classified as being 90° or 180° out of phase. If they are in phase, the coldest air is in the troughs and the warmest air in the ridges. For isotherms to be classified as 180° out of phase, the exact opposite must occur. When 180° out of phase, the waves are fast moving, and their speeds exceed that of the gradient wind within them. When 90° out of phase, the waves move with a speed equal to that of the gradient wind within the wave. In both out-of-phase cases, the temperature advection taking place is significant and the pressure changes are great downstream.

In all the above relationships, the effects of convergence, divergence, and dynamic deepening and filling on temperature are not taken into account. Therefore, these rules are used only to approximate a wave's vector movement.

UPPER-LEVEL HIGHS AND LOWS

In addition to the wave patterns on upper-level charts, closed circulations (lows and highs) also exist. The vertical extent and slope of these systems were discussed in Unit 3, and at this time it might help if you quickly review that material. Basically, these highs and lows are classified according to the vertical variation in their temperature patterns.

- The semipermanent highs (warm core) and lows (cold core) maintain their closed circulations well into the upper troposphere. They are generally stationary with nearly vertical axes.

- Tropical cyclones and thermal lows are classified as warm-core systems, while the major high-pressure systems of Canada and Siberia are cold core. These warm-core systems are very shallow with the exception of mature tropical cyclones, and their closed circulations are confined to the lower half of the troposphere, disappearing with height.

- The dynamic migratory systems, like the semipermanent systems, also retain their closed circulations to appreciable altitudes. The lows slope upward toward the coldest tropospheric air (usually northwest of the surface low), and the highs slope toward the warmest tropospheric air (usually southwest of the surface high). In their final stages of development, those dynamic lows that fully occlude become cold-core with practically vertical axes. The intensification process of dynamic highs is similar to that of the lows, although we don't classify it as an occluding process. It results in a vertically stacked warm-core system.

These dynamic systems also have a relationship with the strongest belt of westerly winds. Dynamic lows are normally poleward of the strongest westerlies, while the dynamic highs are normally equatorward of these winds. Occasionally, this situation gets reversed and a dynamic low ends upon the equatorward side and a high on the poleward side. Highs and lows in this abnormal position are termed CUTOFF CENTERS.

Learning Objective: Define cutoff highs and lows, blocks, zonal and meridional flow, and the jet stream as they relate to upper-air charts.

Cutoff Highs

Cutoff highs occur when warm tropical air is advected poleward into the crest of a steep long-wave ridge. This warm air, and the high that it develops, ends up on the poleward side of the strongest westerlies. In this position, it is "cut off" from receiving additional tropical air.

Cutoff Lows

Cutoff lows take on far more importance than their counterpart highs because of the bad weather that accompanies them. In the Northern Hemisphere, they occur most frequently along the southwest coastal areas of the United States and northwest coastal areas of north Africa. They form when cold polar air is advected equatorward into the base of a deep long-wave trough. This pool of cold air, and the low it develops, ends up on the equatorward side of the strongest westerlies. Figure 8-3-4 shows the typical stages in the development of a cutoff low. The latter stages are often associated with occluding low-pressure systems at the surface; however, cutoff lows are noted for their production of bad weather without frontal or cyclonic circulations at the surface. In the meridional flow of which they are a part, they are the cold valleys between mountains of warm air. On the 500-mb chart, the warm ridge to the west is very sharply oriented southwest to northeast. To the northeast, an intense warm stationary high or ridge BLOCKS the normal path of migratory systems.

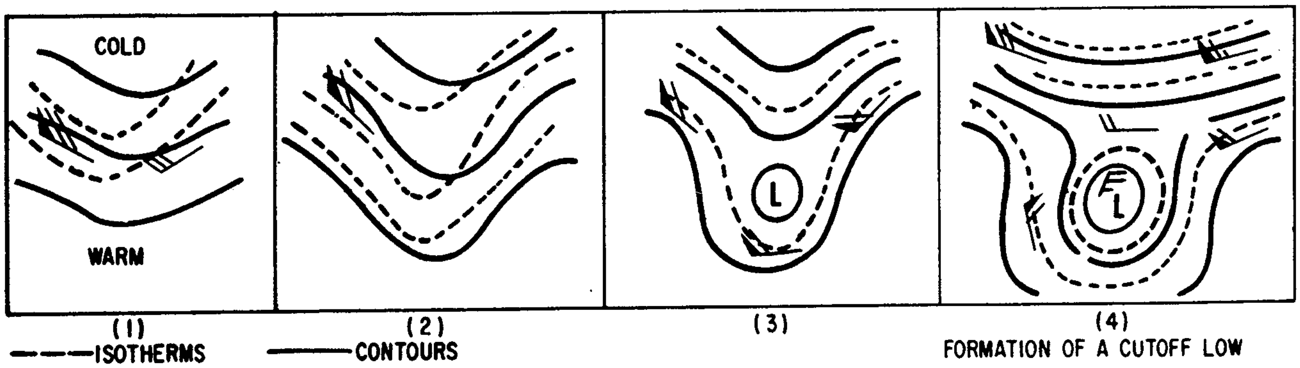

Figure 8-3-4.—Development of cutoff low.

BLOCKS

Blocking is the obstructing of the normal west to east progress of migratory systems. Like cutoffs, blocks are associated with pronounced meridional flow in the upper levels, occur most frequently in spring and least often in autumn, and tend to form in the same geographical regions. In the Northern Hemisphere, they form most frequently in the eastern North Atlantic and eastern North Pacific oceans.

Blocks are warm long-wave ridges or upper-level highs that take up residence in higher latitudes and move very slowly, if at all. As a block sets up and intensifies, it tends to move west (retrograde). Figure 8-3-5 shows three types of

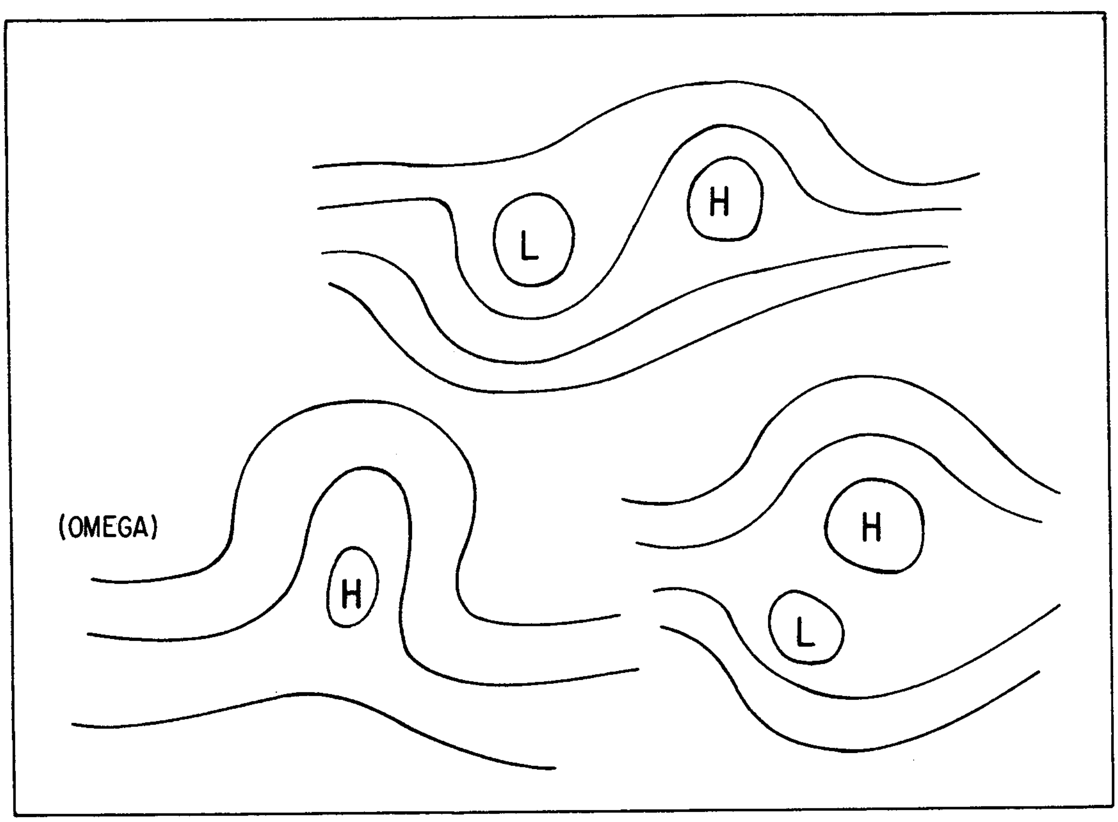

Figure 8-3-5.—Types of blocks.

blocks that occur in the circulation pattern. Blocking highs are either warm-core or dynamic systems, but in either case, they cause a split in the westerlies. The split causes the migratory systems to divert from their normal paths. It is at these times that prolonged periods of good or bad weather occur.

In the winter of 1976-77, a block developed over the western United States. Basically, it lasted from 15 Oct 1976 to 20 Feb 1977. To the west of this block, warm air was transported far north; Alaska basked in balmy weather. Meanwhile to the east of the block, the eastern half of the United States had one of the worst winters on record. Cold arctic air was advected from over the North Pole southward to the Bahamas. The Ohio basin was particularly hard hit and suffered one of its coldest winters on record. The Chesapeake Bay of Virginia and Maryland froze over, and ice breakers were required to keep the shipping lanes open. The citrus industry of Florida suffered severe losses, and the list goes on. In the western half of the United States, drought-like conditions existed, because the block prevented moisture-laden air of the Pacific from reaching the coast. Granted, this block was far from normal, but it serves to show a block's effect on weather patterns.

ZONAL AND MERIDIONAL FLOW

Climatic studies show hemispheric pressure patterns differ in a characteristic fashion depending on whether zonal (east-west) or meridional (north-south) flow dominates within the mid-latitudes. Since winds are pressure generated, a numerical ZONAL INDEX based on horizontal pressure differences between 35° and 55° latitude was developed to measure the strength of these westerly winds. The higher the index, the greater the zonal component. The lower the index, the greater the meridional component.

A particular index may remain nearly stagnant for several weeks, especially in winter, or last for only a few days. Today, zonal indices are seldom computed; merely estimated. The estimate is based on the current circulation and weather patterns. Since reference is made to zonal index, you must have an understanding of what is implied by HIGH, LOW, and CHANGING zonal indices.

High Zonal Index

With a high zonal index, long-wave ridges and troughs are weaker (have less amplitude) and fewer in number. The Icelandic and Aleutian lows are well developed and are at, or slightly north of, their normal positions. Their axes and the orientation of their associated troughs are east-west. The Atlantic and Pacific subtropical highs are north of their normal positions, and their orientation is strongly east-west. The Great Basin and Siberian highs are present. There's little westward extension to the high over the Great Basin. (The Great Basin encompasses all of Nevada, the western one-third of Utah, and portions of eastern California, southern Oregon and Idaho.) Highs are absent at higher latitudes. Fronts are well north and predominantly oriented east-west. They and their associated lows move rapidly east ward. The highs of mid-latitudes are moderately developed and also progress east at a rapid pace. There are fewer polar outbreaks; therefore, the polar regions get colder, while mid-latitude temperatures are moderate. The weather is generally fair in the mid-latitudes. The stormiest weather is along 60°N.

Low Zonal Index

With a low zonal index, long-wave ridges and troughs are very steep (have great amplitude), and there are more of them. Cutoff centers are common. Short waves have larger amplitudes than normal and are a predominant feature of the 700-mb chart. The Icelandic and Aleutian lows are weak, are split into two cells, and are oriented north-south. The Atlantic and Pacific subtropical highs are weak, split, oriented north-south, and are centered farther south than normal. The polar highs merge and are strongly developed. Frontal systems are more common, with sharp temperature contrasts across them. The fronts are oriented north-south. Stormy weather is frequent in low latitudes, while high latitudes experience storm-free weather and mild temperatures.

Changing Zonal Index

Hemispheric weather patterns are, for the most part, always changing. As the pattern changes, so does the zonal index. A pattern change from a high-index situation to a low, or vice versa, is simply one of transition. It doesn't happen overnight. A changing index is one that is either increasing or decreasing. INCREASING or DECREASING indices are also based on the circulation and weather pattern.

INCREASING ZONAL INDEX.— With an increasing zonal index, the number of long waves in the weather pattern decreases. Cutoff centers weaken. Migratory systems intensify and speed up, especially in higher latitudes, resulting in a more east-west orientation of associated frontal systems. The eastern cells of the Icelandic and Aleutian lows and subtropical highs weaken while moving eastward. The western cells of these systems move northeastward toward their normal positions. A polar high stagnates over the Great Basin.

DECREASING ZONAL INDEX.— With a decreasing zonal index, the number of long waves in the weather pattern increases. Cutoff centers are more likely to develop. Migratory systems slow down, especially in the higher latitudes. Frontal systems gradually shift from a east-west orientation to one that is more north-south. The Aleutian and Icelandic lows and subtropical highs move southward and begin to split. Polar highs intensify, and outbreaks of cold polar air into lower latitudes take place.

JET STREAMS

Winds in the upper-level circulation pattern reach maximum speeds in narrow streams that meander in wavelike fashion above both hemispheres. By definition, when these streams of high-speed winds are thousands of miles long, hundreds of miles wide, and a few miles deep (vertical extent), they are termed jet streams.

Jet streams are three-dimensional features. They have length, width and depth. Isotach analyses done on upper-level constant-pressure charts provide us with a view of two of the dimensions. They outline a jet stream's axis and show its width latitudinally and length horizontally. These analyses also show that jets are not continuous around the globe; they stop, start, split, merge, and can exist side by side within a few hundred miles of one another or be thousands of miles apart. The third dimension (depth) is best determined from vertical wind profiles developed from wind information obtained from radiosonde, rawinsonde, and pilot reports. However, you can obtain a crude picture of a jet's depth by stacking upper-level charts. Three-dimensional studies also reveal a core of higher speed winds existing within each jet stream. The winds in the jet core differ from those elsewhere in the stream in that they are stronger and have a constant velocity. In other words, their direction and speed are unchanging. The importance of this is related to wind shear, experienced in and around jet streams. If there is no change in wind speed and direction in the core, there can be no wind shear there. Outside of the core, however, wind shear can be significant. The shear associated with the polar-front jet stream will be discussed in depth later in this unit.

Jet Streams of the World

When jet stream winds were first discovered, they were thought to be part of a solitary stream extending around the globe over the mid-latitudes. Research proved this not to be the case, and as other jet streams were discovered, they were named. Their names are derived from the regions over which the jet streams occur and, in some cases, their association with other features. The jet streams covered in this unit are as follows:

- Polar-front or mid-latitude jet stream
- Subtropical jet stream
- Tropical easterly jet stream
- Polar-night jet stream

Most studies center around the middle-latitude jet streams of the Northern Hemisphere, because the availability of upper-level data in this region far exceeds that of other areas of the world, and because these are the jet streams that impact the majority of the world's population. With this in mind, most of our discussion will focus on the polar-front and subtropical jet streams.

POLAR-FRONT JET STREAMS.— These jet streams are found in both hemispheres and are associated with the principal frontal zones and low-pressure systems of the middle and subpolar latitudes. They appear as single or multiple systems on upper-level charts. A single polar-front jet occurs characteristically during high index cycles, when there are few long waves and the westerlies are predominantly zonal. Multiple jets occur during low index cycles, when there are numerous long waves and the westerlies are predominantly meridional.

The Single Polar-front Jet.— When the upper-level westerlies are well organized, a single polar-front jet exists. The stream lies west to east, with cold air north of the axis and warm air south. The wind along the axis shows areas of alternating maximum and minimum wind speeds, known as jet maxima and jet minima. Each maximum is associated with a short-wave trough and moves with a speed proportional to the short-wave's speed at the 700-mb level. The wind speed in these maxima may easily exceed 250 knots, and the distance between each maximum varies from about 10° to 25° longitude. The difference in wind speed between the maxima and minima can be as much as 100 knots. As the westerlies begin to shift from zonal to meridional, the single polar-front jet often splits into two or more branches. We are then faced with multiple polar-front jet streams.

Multiple Polar-front Jets.—When the upper-level westerlies become predominantly meridional, outbreaks of polar and/or arctic air become common. A jet accompanies each outbreak and is directly associated with the frontal zone. On the average, each jet drifts equatorward at a rate of 30 nautical miles per day. They intensify over the mid-latitudes, but weaken as their associated polar or arctic air moderates in the lower latitudes. This equatorward migration is not regular, because some jets become stationary or even drift back to the north. The wind along their axes is more or less uniform, and the jet maxima and minima disappear. Split jet streams are common, and the most important are those associated with the formation of blocks in the upper-level circulation. Numerous small offshoots of jet flow, known as jet fingers, are also common. They are weaker than the primary jet, and at times the fingers may be separated by as little as 5° latitude. Figure 8-3-6 illustrates these features. Multiple jets usually retain their identity around the hemisphere, but occasionally, they merge. When jets merge, the wind speed in the stream intensifies. As the circulation pattern changes, the westerlies once again organize into strong zonal flow, and a single polar-front jet is reestablished.

From the above information, you can see that on any given day, the number of polar-front jets, their strength, and location may vary. However, climatologists portray them as a single feature on climatic charts and provide mean statistics about them. In the absence of real-time data, climatic information becomes very important.

CLIMATOLOGY.— Climatic charts depicting jet stream winds are somewhat misleading. They show a single jet (figures 8-3-7 and 8-3-8), when

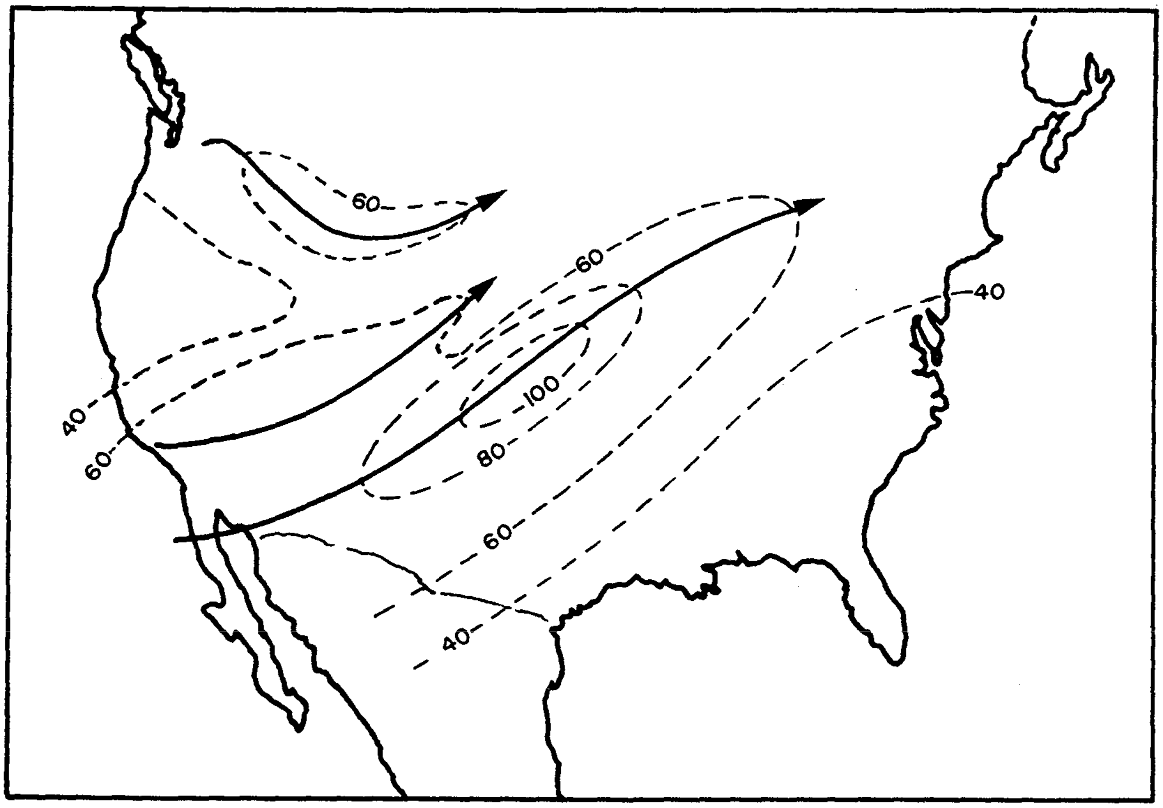

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Figure 8-3-6.—Jet fingers.

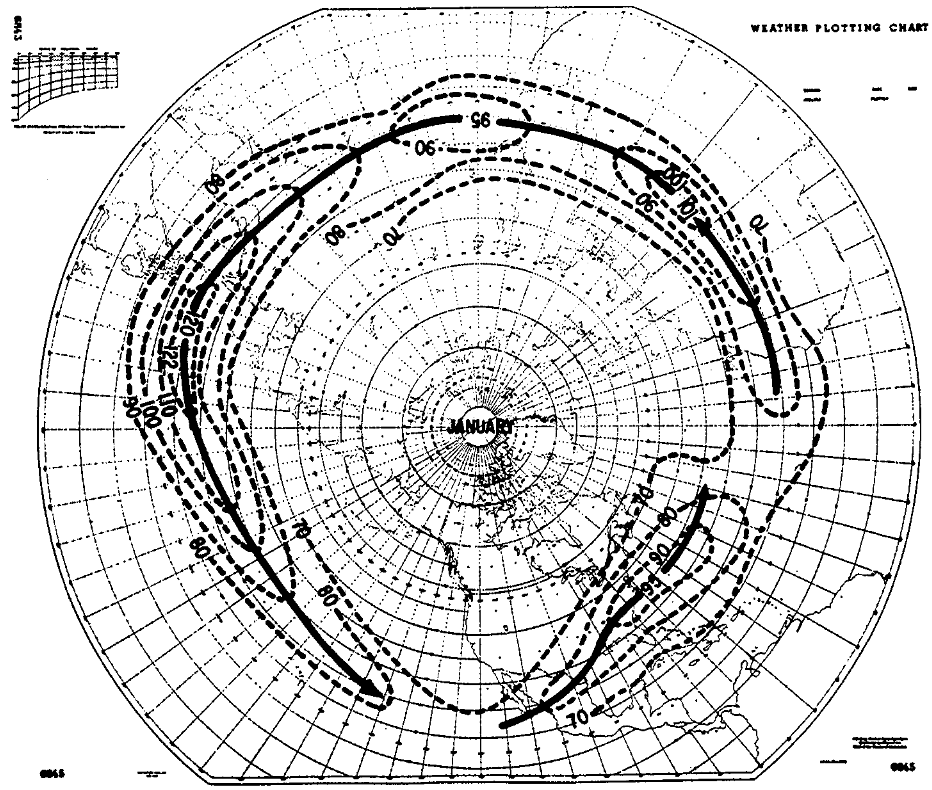

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Figure 8-3-7.—Mean seasonal jet stream distribution for January.

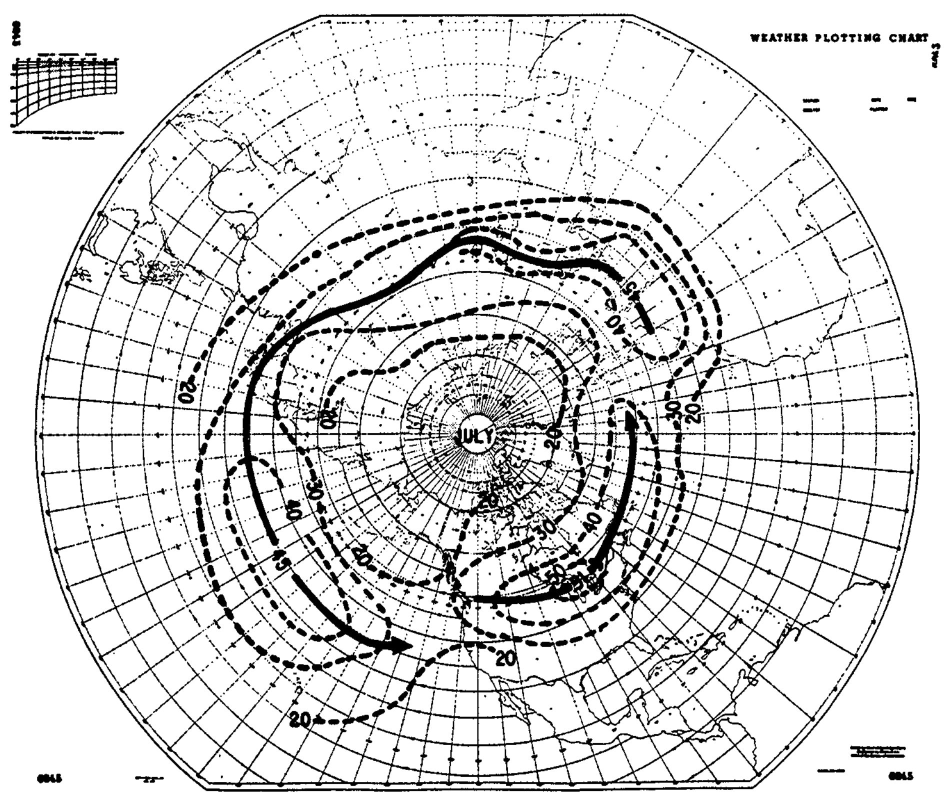

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Figure 8-3-8.—Mean seasonal jet stream distribution for July.

in fact there are more than one. This single or mean jet stream is the result of the averaging of all the upper-level westerlies over a given hemisphere for a specific period of time.

The mean jet stream of the Northern Hemisphere is found between 25°N and 45°N, depending on the season of the year. The polar-front jet and subtropical jet both exist within these boundaries; however, the mean should not be thought of as being specifically representative of either. REMEMBER, THE MEAN JET IS DERIVED FROM ALL UPPER-LEVEL WESTERLIES OVER A GIVEN HEMISPHERE, FOR A GIVEN PERIOD.

Climatic statistics show that the mean jet of winter is stronger (70 knots) than the mean of summer (35 knots), that the jet core is slightly higher in winter, and that the depth of jet winds is much greater in winter. This last fact is the reason we use the 300-mb chart in jet stream analysis during the winter and the 200-mb chart during the summer.

No matter what the season, certain areas of the Northern Hemisphere have greater jet stream winds than others. The strongest jets are encountered over Japan when a polar-front jet and a subtropical jet merge over the area. Winds close to 300 knots are not uncommon when this merger takes place. Other areas where strong jets are the norm are off the east coast of the United States, over the Sahara Desert, and the Arabian Sea. From this, we can say that higher velocity jet streams are concentrated near east coasts of continents, and lower velocity jets are found over west coasts of continents. There is also a greater latitudinal fluctuation in jet streams over the west coasts of continents.

Polar-front Jet Stream Relationships

Researchers have associated many meteorological features with the jet streams of the mid-latitudes in order to further our understanding of these high-speed winds and their effect on weather. Obviously, they're related to polar fronts, but they are also related to migratory low-pressure systems, various weather conditions, the tropopause, clouds, and turbulence. Lets begin with the jet's relationship to fronts.

FRONTAL RELATIONSHIP.— The jet stream's location relative to its front is dependent on the height the front attains as it slopes up from the surface.

1. If the front reaches (intersects) the 500-mb level over a station, the jet core lies almost directly above the station.
2. If the front fails to reach the 500-mb level over a station, the jet core lies north of the station.
3, If the front intersects a level above the 500-mb level over a station, the jet core lies south of the station.

Frontal intersections with constant-pressure levels show up as isotherm packing at each level the front intersects. If more than one zone of packing exists on a chart, more than one front extends to that level. The strongest jet is the one associated with the zone of greatest packing (strongest temperature gradient) at the 500-mb level or below. This packing may be as great as 10°C in 45 miles, but usually ranges between 10°C in 90 miles and 10°C in 150 miles. The packing of isotherms is weaker through ridges and stronger through troughs, and the width of the packing, as represented on the 500-mb chart, equates to the width of the jet core above it.

POLAR FRONT AND DEVELOPING LOWS.— A polar-front jet stream is associated with surface low-pressure systems in a very simple way. As a surface frontal wave develops, a surface low is formed. The jet moves southward, pushing cold air into the west side of the low, and the low intensifies. When the polar front occludes, the jet moves south of the low center and crosses the polar front at the apex (triple point) of the occlusion. In other words, it parallels the polar front and remains north of the migratory surface lows until after the polar front occludes. The following is a list of jet stream relationships to fronts and surface lows:

1. The jet stream remains north of unoccluded lows.
2. In a series of migratory lows (cyclone family), each low is associated with a jet maximum. But remember, every jet maximum is not necessarily associated with a surface low.

The two most common positions for these lows and their jet maxima are given in figure 8-3-9. NOTE: As the upstream low deepens, a jet maximum develops along the axis to the west of the low. This cycle of development continues with each developing frontal wave.

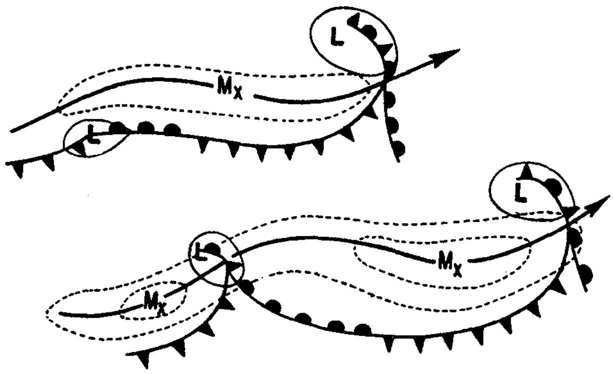

Figure 8-3-9.—Usual position of surface lows in relation to moving jet maxima.

3. The jet stream parallels the warm-sector isobars of surface lows.
4. The jet stream is found south of occluded lows, near the point of occlusion.
5. The jet stream is perpendicular to occlusions.
6. The jet stream roughly parallels the isobars south of a cold, slow-moving surface low.

HIGHS.— The polar-front jet is also related to surface high-pressure systems.

1. The jet stream roughly parallels the isobars north of a warm, slow-moving high.
2. When a cold, migratory polar high stagnates and begins to warm up, the jet stream configuration upstream can change drastically. If a cold-core high becomes warm-cored, the original jet dissipates and a new jet forms to the north.

WEATHER.— The weather associated with polar-front jet streams varies. However, bad weather is more often associated with them than good, especially in winter. Extensive bad weather is normally found between the surface warm front of a developing or mature surface low and the jet axis to the north. Also, when a cold front has a very shallow slope and the jet is positioned well back in the cold air (usually 600 miles), the cold front will be overrun by a warm southwest flow of air. This overrunning produces extensive cloudiness and continuous precipitation. The occurrence of precipitation associated with the jet stream is controlled primarily by the distribution of wind shear and curvature along the jet stream. There is general agreement that the highest incidence of precipitation almost straddles the jet axis, with a slight bias toward the cold-air side.

Severe frontal thunderstorms are also thought to be related to the jet stream. These thunderstorms form in regions having strong vertical wind shears, which are commonly found beneath a jet. The energy of the jet is transferred from the large-scale upper-level circulation to the smaller scale cyclonic circulation at the surface (the surface low). From here, the energy is transferred to an even smaller scale circulation, the thunderstorm. If a tornado was to develop from the thunderstorm, the energy transfer would have been carried one step further. This type of energy transfer is known as CONSERVATION OF ANGULAR MOMENTUM.

THE TROPOPAUSE.— The mid-latitude or polar-front jet streams are associated with breaks in the tropopause. North of a jet the tropopause is low, and south of the jet it's high. These jets can exist without a break in the tropopause, and when this happens, the tropopause shows a very steep slope in the region of maximum wind. There will also be times when two tropopauses overlap, and at these times, the jet is often found between them. The mid-latitude or polar-front jets are associated with the break between the subarctic and mid-latitude tropopauses (30° to 50° lat.).

CLOUDS.— Cirrus is the cloud most often associated with jet streams. It usually forms on the warm-air side of the axis where lateral shear is relatively weak and extends to within several thousand feet of the tropopause. Occasionally, a sharp discontinuity in the cirrus takes place near the jet core; the skies become cloudless immediately north of the core. If cirrus extends north of the core, there is usually a narrow break in the cirrus at the core itself. Most of the time there are no clouds at or above the level of the jet core. On the average, jet stream cirrus is 500 feet thick, but can range from a hundred feet to several thousand feet.

Four cloud patterns are associated with the mid-latitude and polar-front jets, and at least three of the four must be present for a jet to exist. They often completely cover the sky and have bands at right angles to the air flow. The cloud patterns are as follows:

1. Lines of cirrus in bands (H4, 5, and 6)
2. Patches of cirrocumulus (H9) or altocumulus castellanus (M8)
3. Lenticular clouds in waves (M4 or M7)
4. Bands of altocumulus (M3 or M5)

Because bad weather often accompanies polar-front jets, lower cloud shields will often obscure the higher clouds. Your ability to recognize a jet stream's existence by cloud patterns will be dependent on your seeing the clouds. With pilots it's a different story. Flying above the bad weather, they'll get a first hand look at the higher clouds. It's important that they be able to recognize the clouds associated with jet streams, in order to warn passengers and crew of possible impending turbulence. The turbulence that accompanies jet streams is known as clear-air turbulence. Don't let the name fool you; this type of turbulence occurs both inside and outside clouds.

CLEAR-AIR TURBULENCE.— Pilots expect turbulence in clouds, but they are not always expecting it in cloud-free skies. Pre-flight briefings, which include jet stream locations and altitudes in relation to flight paths and altitudes, give pilots advanced warning as to when and where they may experience clear-air turbulence.

The wind shear within the jet stream causes this rough cobblestone-type of bumpiness. Clear-air turbulence buffets the crew, passengers, and equipment, and because of the stresses it places on the airframe, it decreases an aircraft's useful life.

Horizontal Wind Shear.— Along the axis of a jet stream, wind speeds decrease very rapidly on the cold-air side of the core and more slowly on the warm-air side. Therefore, horizontal wind shear is much greater on the cold-air side of jet streams. For example, the wind speeds may drop off 100 knots within 100 miles on the cold-air side of the core, while only dropping off 25 knots over the same distance on the warm side. Figure 8-3-10 illustrates the packing of isotachs on the cold-air side of the jet axis.

Vertical Wind Shear.— The wind speeds above and below the core can also decrease very rapidly over short distances. A decrease of 30 to 40 knots in 1,000 feet is not uncommon. Normally, the shear above the core is much greater than that below the core. However, when the jet core is above a frontal zone, the shear beneath the core exceeds the shear above the core. Figure 8-3-11

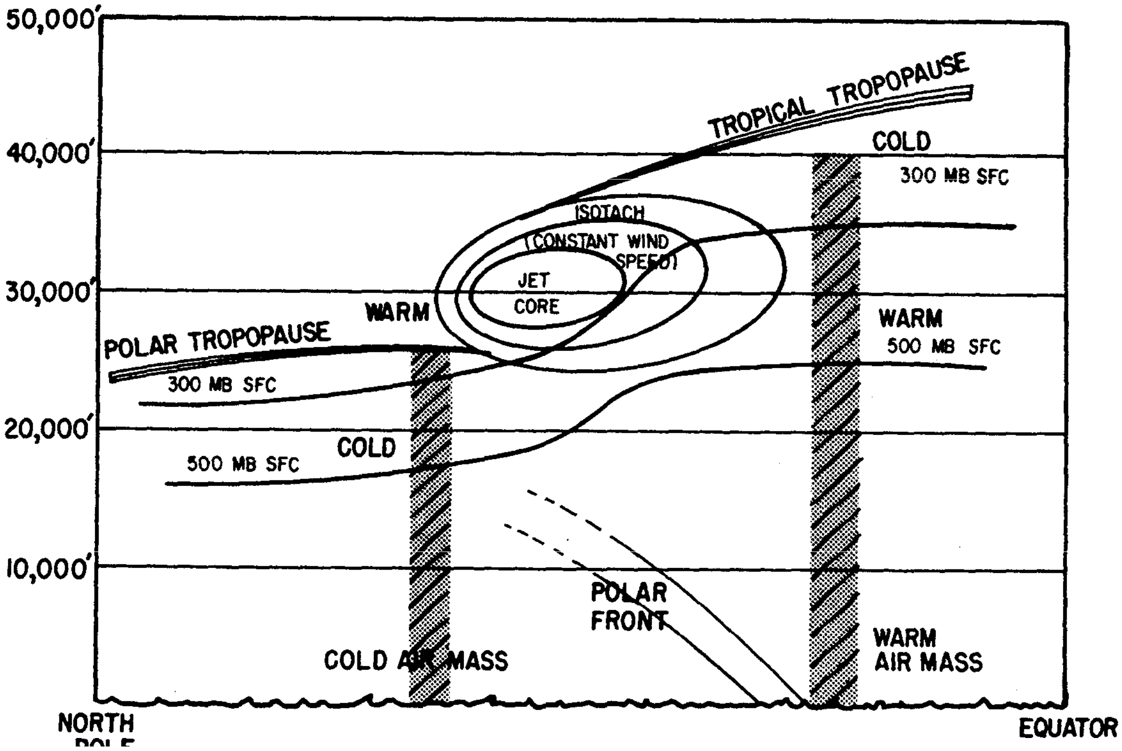

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Figure 8-3-10.—Vertical cross section of a model of the jet stream—winds and fronts.

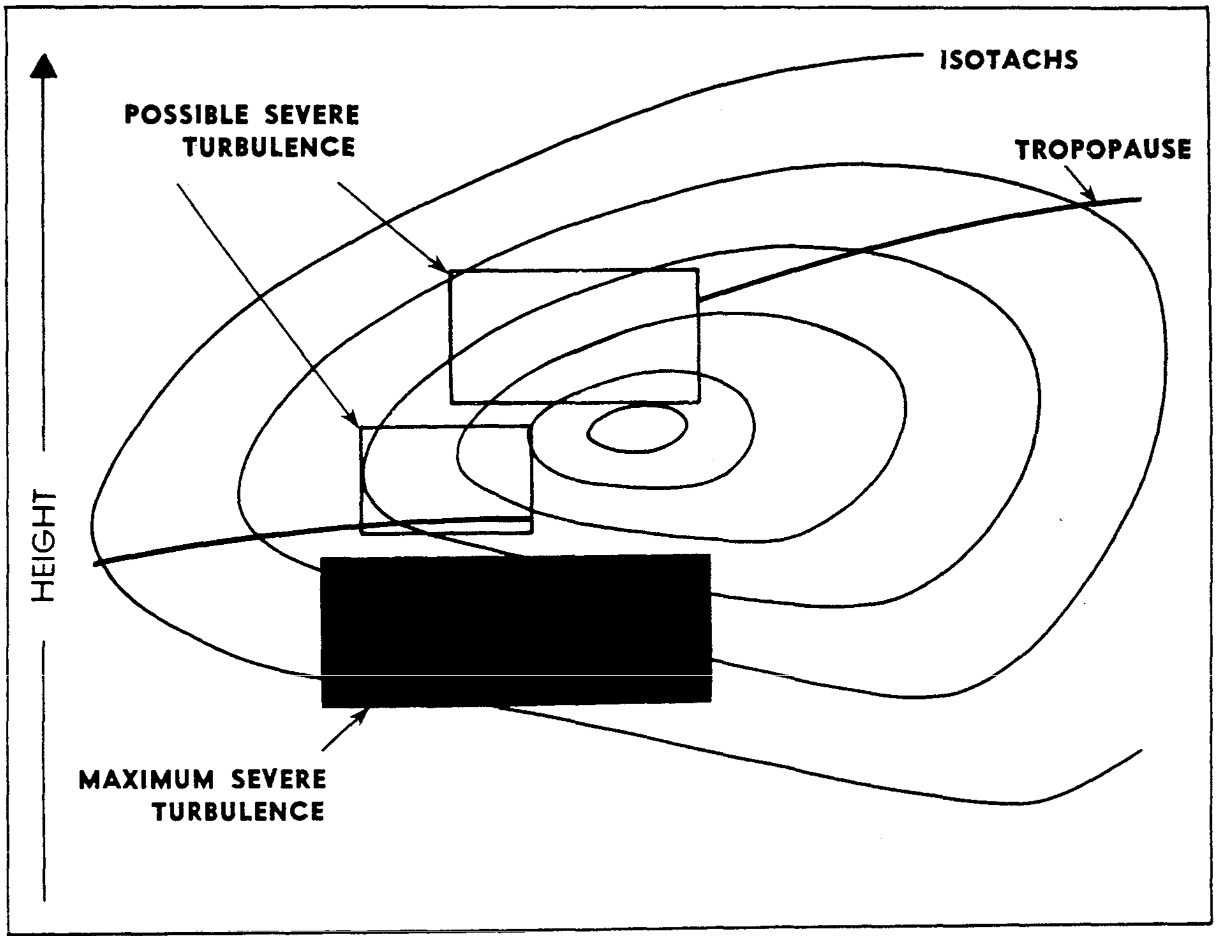

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Figure 8-3-11.—Probable areas of clear-air turbulence.

illustrates the most probable areas for aircraft to encounter clear-air turbulence. The two most probable locations are just below and to the left of the core between 22,000 and 28,000 feet and directly above the core in the vicinity of the tropopause between 35,000 and 50,000 feet.

Even though there's the possibility of encountering clear air turbulence, many flights take advantage of these jet stream currents. Flying with the jet saves time and money. The savings in fuel dollars can be considerable. On the other hand, going against the jet requires more fuel, possibly more stops enroute, and additional flight planning.

SUBTROPICAL JET STREAMS.— These jets, like the polar-front jets, are best developed in winter and early spring. During summer, in the Northern Hemisphere, the subtropical jet weakens considerably, and it is only identifiable in sporadic velocity streaks around the globe. During winter, subtropical jets intensify and can be found between 20° and 50° latitude. Their maximum speed approaches 300 knots, although these higher wind speeds are associated with their merger with polar-front jets. The core is most frequently found between 35,000 and 40,000 feet. A subsidence motion accompanies subtropical jets and gives rise to predominantly fair weather in areas they pass over. These jets are also remarkably persistent from time to time, but they do fluctuate daily. Sometimes they drift northward and merge with a polar-front jet. Over Asia in summer, the subtropical jet is replaced by the tropical easterly jet stream.

TROPICAL EASTERLY JET STREAM.— This jet occurs near the tropopause over Southeast Asia, India, and Africa during summer. The strongest winds are over southern India, but they are not as intense as the winds encountered in polar-front or subtropical jet streams. This jet is closely connected to the Indian and African summer monsoons. The existence of this jet implies that there is a deep layer of warm air to the north of the jet and colder air to the south over the Indian Ocean. This warm air is of course associated with the maximum heating taking place over India in summer, while the colder air is over the ocean. The difference in heating and cooling and the ensuing pressure gradient is what drives this jet.

POLAR-NIGHT JET STREAM.— This jet meanders through the upper stratosphere over the poles. It occurs only during the long winter night. Remember, night is 6 months long over the pole in which winter is occurring. The polar stratosphere undergoes appreciable cooling due to the lack of solar radiation. The horizontal temperature gradient is strongly established between the equator and the pole, and the pressure gradient creates this westerly jet. The temperature gradient breaks down intermittently during middle and late winter in the Northern Hemisphere; therefore, the jet is intermittent at these times. In the Southern Hemisphere the temperature gradient and jet disappear rather abruptly near the time of the spring equinox.

UNIT 8—LESSON 4

CONVERGENCE AND DIVERGENCE

OVERVIEW

Define convergence and divergence and describe the importance of each in the science of meteorology.

OUTLINE

CONVERGENCE AND DIVERGENCE

Convergence is the accumulation of air in a region or layer of the atmosphere, while *divergence* is the depletion of air in a region or layer. The layer of maximum convergence and divergence occurs between the 300- and 200-mb levels. Coincidently, this is also the layer of maximum winds in the atmosphere; cores of jet streams are usually found here. These high-speed winds are directly related to convergence and divergence. The combined effect of wind direction and speed (velocity) is what produces convergent and divergent air flow.

Learning Objective: Define convergence and divergence, and describe the importance of each in the science of meteorology.

IMPORTANCE OF CONVERGENCE AND DIVERGENCE

The importance of convergence and divergence is related to pressure changes at the surface and height changes of the constant-pressure levels. As the air accumulates in the 300-200-mb stratum over a region, greater pressure is exerted throughout the atmosphere. It's like inflating a tire.

As the air flows into the stratum, the pressure increases. Barometric pressure at the surface rises, as do the heights of the constant-pressure levels. Put another way, we say that upper-level convergence causes pressure and height rises. The exact opposite effect takes place when air is depleted from the 300-200-mb stratum. This is where we deflate the tire. As the air flows out of this stratum, pressure is lost. Barometric pressure at the surface falls, as do the heights of the constant-pressure levels. We say that upper-level divergence causes pressure and height falls.

Convergence and divergence aren't the only processes at work in the atmosphere that can cause pressure and height changes, but you're going to hear these terms with increasing regularity as you progress up the AG rate ladder. They are primarily used by forecasters to explain why systems are expected to fill or deepen during the forecast period. Since the upper winds are the producer of convergence and divergence, you should be able to recognize the flow patterns associated with their production. The flow patterns range from simple to complex.

CONVERGENCE AND DIVERGENCE (SIMPLE MOTIONS)

In order for convergence to take place, the winds must be such as to result in a net inflow of air into a layer or region. At the surface, low-pressure systems are associated with convergent flow. The winds cross isobars toward the center of the low and push the air in the center upward into the atmosphere. The currents are illustrated in figure 8-4-1. The upward vertical motion is a prime contributor to the occurrence of precipitation. In meteorology, convergence is classified as horizontal or vertical, because there are horizontal and vertical currents occurring in the atmosphere.

In order for divergence to take place, the winds must be such as to result in a net outflow of air from a layer or region. High-pressure systems are associated with divergent flow. The winds cross isobars, flowing out from the high's center and depleting the air within the high. The air above the high sinks to replace the outflow of air at the surface. This downward vertical motion (subsidence) is associated with dry air. Divergence can also be classified as horizontal or vertical, depending on the wind's axis. See figure 8-4-1.

The simplest form of convergence and divergence is the type that results from wind direction alone. Two flows of air brought together, no matter what the angle, result in convergence. Where the air flow splits and winds go in different directions, divergence is occurring. Figure 8-4-2 illustrates these types of convergent and divergent air flow.

Wind speed in relation to wind direction is also a contributor to convergence and divergence. If the wind speed decreases downstream, there's a net inflow of air into the region, and convergence takes place. If wind speeds increase downstream, there's a net outflow of air from the region, and divergence occurs. In an area of uniform wind speeds, if the winds fan out (split), divergence occurs. If these same winds are brought together, convergence occurs. See figure 8-4-2, view B.

The fact that contours converge or diverge doesn't necessarily indicate convergence or divergence, because wind speeds must also be considered. If wind speeds increase downstream and the contours spread apart, supergradient winds are said to be occurring. This combination of wind direction and speed produces divergence. On the other hand, if wind speeds decrease downstream and the contours converge,

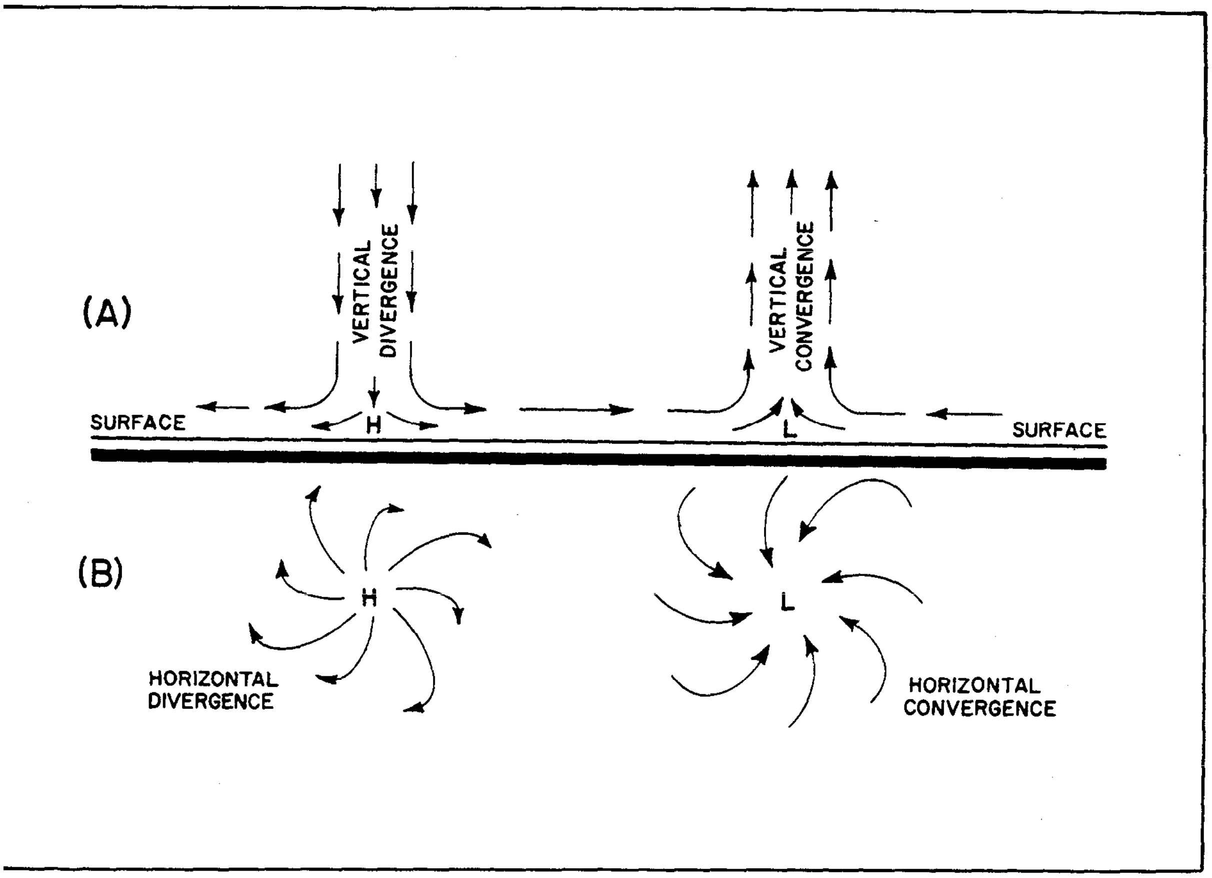

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Figure 8-4-1.—Convergence and divergence. (A) Vertical perspective; (B) Horizontal perspective (Northern Hemisphere).

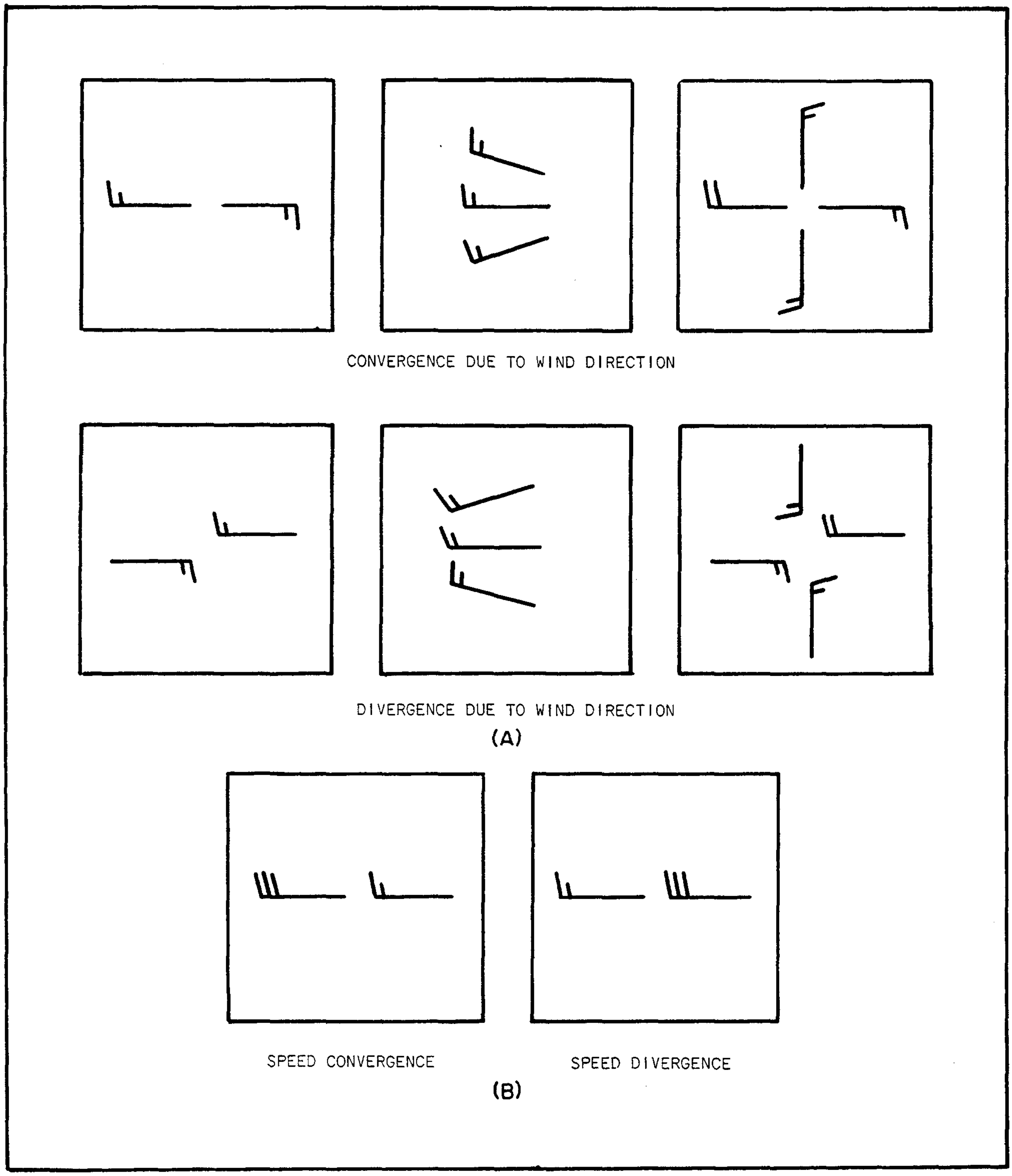

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Figure 8-4-2.—Convergence and divergence; (A) directional, (B) speed.

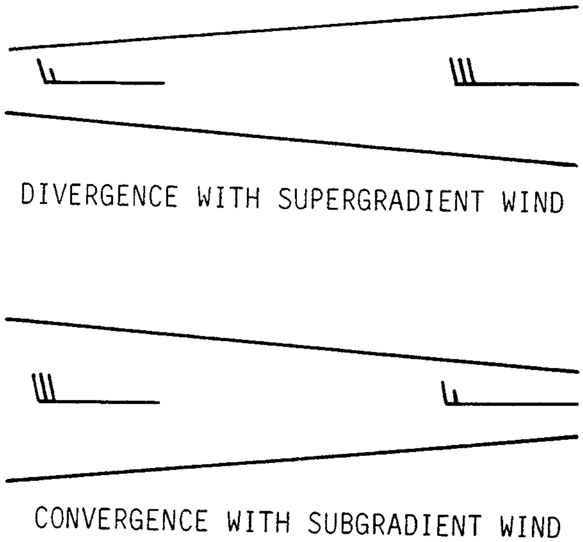

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Figure 8-4-3.—Convergence and divergence in supergradient and subgradient air flow.

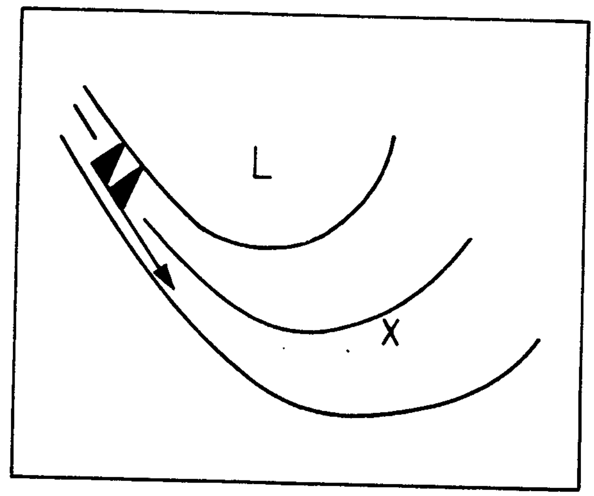

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Figure 8-4-4.—Divergence illustrated.

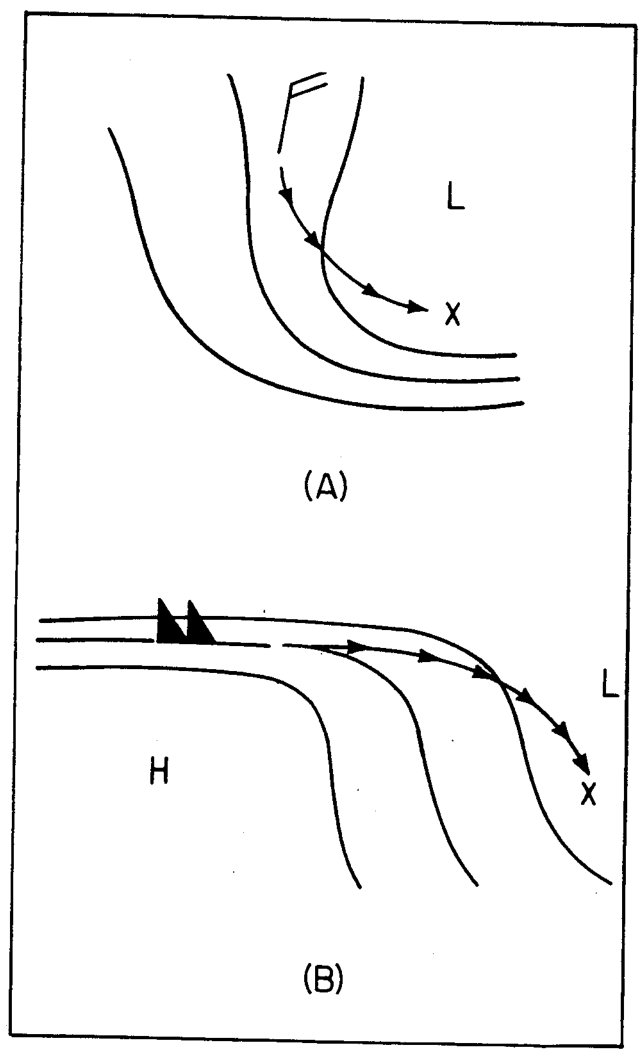

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Figure 8-4-5.—Convergence illustrated,

subgradient winds are occurring. This combination produces convergence. See figure 8-4-3.

There are other cases where it is difficult to tell whether divergence or convergence is taking place. When wind speed decreases downstream and the contours spread apart, both convergence and divergence are indicated. The wind speed suggests convergence, but the spreading contours suggest divergence. A similar situation arises when wind speeds decrease downstream and the contours converge. Here, we're looking at speed divergence and directional convergence. These are complex motions, however, and a special evaluation is required to determine the net inflow or outflow.

CONVERGENCE AND DIVERGENCE (COMPLEX MOTIONS)

The 300-mb chart is the primary chart used to determine areas of horizontal convergence and divergence; however, if a sparsity of reports limits its use, the 500-mb chart may be substituted.

Looking at an analyzed 300-mb chart, you have to examine the winds in relation to height contours to locate areas of divergence. Look for areas where high-speed winds are approaching weak cyclonically curved contour gradients. A jet maximum moving south on the west side of a major trough is a good example of this type of wind-contour relationship. See figure 8-4-4. The jet winds push air at excessive speeds, and when the air is forced to negotiate the curve in the contour (much like a car trying to negotiate a bend in the road while moving at high speed), it fails. The air continues in a straight line and crosses the contours to the right of its intended path. Based on resultant forces, where the contours weaken and curve cyclonically, the Coriolis force and centrifugal force combine to overpower the pressure gradient force; and the air flows out of the system, across contours, into an area of higher heights. Divergence occurs downstream to the left of the path of the outflowing air. If these high-speed winds continue for a sustained period of time, large height falls occur in this area.

Areas of convergence are associated with two wind-contour relationships. The first is where low-speed winds approach strong cyclonically curved contour gradients (fig. 8-4-5, view A). A good example of this is where a jet maximum is located on the east side of a major trough and weak winds are on the west side. As the air on the west side moves slowly toward the base of the trough, it encounters the stronger cyclonically curved contour gradient associated with the trailing edge of jet stream winds. The pressure gradient becomes very intense in this area and the Coriolis and centrifugal forces are unable to balance it. In order to reestablish the balance of forces, the slow-moving air is forced to move to the left of its intended path. As the air moves to the left, it crosses contours, and there's a net inflow of air into the trough. The convergence occurs where the air flows across the contours.

The second wind-contour relationship that creates convergence occurs when high-speed winds approach winds of lower speeds in weak anti-cyclonically curved contour gradients (fig. 8-4-5, view B). An example you might look for would have a jet maximum approaching the top of a sharply curved ridge. As the air is pushed into the ridge at excessive speeds, it is unable to make the sharp turn necessary to follow the contours. Here the centrifugal force and pressure gradient force combine to overpower Coriolis force. The air is forced to cross the contours to the left of its intended path and flows into the downstream trough. The convergence occurs to the right of the path of inflowing air.

UNIT 8—LESSON 5

ROTATIONAL MOTION AS IT AFFECTS THE ATMOSPHERE

<u>OVERVIEW</u>

Define vorticity and explain the two types.

State how CAVT tables are used.

<u>OUTLINE</u>

Vorticity

Constant absolute vorticity trajectories (CAVT)

ROTATIONAL MOTION

An air-parcel pushing through Earth's atmosphere has three spinning motions: (1) It rotates about its own axis; (2) it rotates about the axis of its pressure system; and (3) it rotates about Earth. All of these motions occur simultaneously. Measuring the spin imparted to an air parcel is known as vorticity measurement. In meteorology, vorticity measurements are applied only to those spinning motions having axes perpendicular to Earth's surface. Air spinning about a high- or low-pressure system is an example of such motion, as these systems extend vertically upward into the atmosphere (a perpendicular axis). Vorticity measurement gained prominence in the 1950's and 1960's, and over the last 30 years it has become an important forecasting tool, especially in the field of numerical forecasting. This discussion will be brief, as I will simply explain the three spinning motions and the two types of vorticity. A more in-depth discussion of this subject and its forecasting application can be found in the *AG 1* rate training manual.

Learning Objective: Define vorticity and explain the two types.

VORTICITY

A parcel has vorticity when it spins as it moves along its path. Depending on the direction of the spin, the vorticity is either positive or negative. In the Northern Hemisphere, if a parcel spins in a counterclockwise direction (cyclonically), it has positive vorticity. If a parcel spins clockwise (anticyclonically), it has negative vorticity. If a parcel does not spin, it is said to have zero vorticity.

For an example of how spin is imparted to air parcels, we could drop a chip of wood into a stream and watch its progress. The chip will move downstream with the flow of water, but it may or may not spin. If it spins, it has vorticity. Two properties of the stream cause the chip to spin: (1) Water moving faster on one side of the chip than on the other (a shear effect) and (2) curves in the stream bed (a curvature effect). When we measure the spin created by these two components (shear and curvature), we can determine relative vorticity.

Relative Vorticity

As we have just seen in our example, relative vorticity is observable. You can examine any point on an upper-air chart and say whether an air parcel at that point does or does not have vorticity. It's simply a matter of checking the wind on either side of the point to determine the shear and noting the parcel's position in the air stream for the effect of curvature. Relative vorticity is a measure of the spin created by shear and by curvature.

SHEAR.— Let's examine the shear effect by looking at small air parcels in an upper-air pattern of straight contours. The wind shear in our

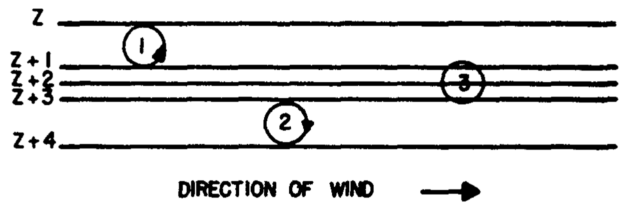

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Figure 8-5-1.—Illustration of vorticity due to the shear effect.

example causes two of the three parcels to rotate. See figure 8-5-1.

1. Parcel No. 1 has stronger wind speeds to its right. As the parcel moves along, it is rotated in a counterclockwise direction and thus has positive vorticity.

2. Parcel No. 2 has the stronger speed to its left; therefore, it rotates in a clockwise direction as it moves along. Its vorticity is negative.

3. Parcel No. 3 has speeds evenly distributed. There's no shear. The parcel moves, but it does not rotate. It has zero vorticity.

Remember, air parcels have vorticity (rotation) when the wind speed is stronger on one side of the parcel than on the other.

CURVATURE.— Vorticity also results from curvature of the air flow or path. Examine figure 8-5-2 and try to imagine the progress (and direction of spin) of an air parcel traveling in the stream. In the trough and ridge, the diameter of the parcel is rotated from the solid line to the dotted position (because of the northerly and southerly wind components on either side of the trough and ridge lines). Parcels have

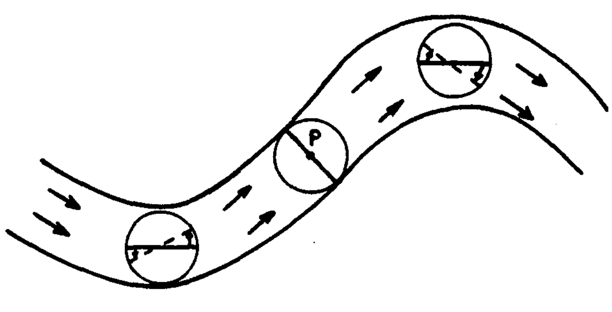

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Figure 8-5-2.—Illustration of vorticity due to curvature effect.

counterclockwise rotation in troughs (positive vorticity) and clockwise rotation (negative vorticity) in ridges. At a point between a trough and a ridge where there is no curvature, the inflection point, there is no spin imparted. This is shown at point P in figure 8-5-2.

COMBINED EFFECTS.— To find the relative vorticity of a given parcel, you must consider both shear and curvature. It is quite possible that the two effects will counteract each other. That is, shear may indicate positive vorticity, and curvature indicate negative vorticity, or vice versa. See figure 8-5-3.

When shear and curvature counteract each other, the amount of vorticity in each must be measured. The two figures are then added together algebraically to determine if vorticity is positive or negative. Measurements also indicate whether the vorticity is increasing or decreasing. When vorticity is increasing, the rotation is becoming more cyclonic. When it decreases, the converse is true.

Absolute Vorticity

To this point we have discussed two of the factors that control the spin of air parcels. The third factor is Earth's rotation. The effect of this rotation is seen only from space. There, we can see the effect of Coriolis force. An air parcel traveling from the equator toward the poles has its trajectory altered because Earth is rotating beneath the parcel. A parcel moving north in the Northern Hemisphere appears to curve to the right. Since it curves, a spin is imparted to it. This is known as the Coriolis factor or parameter. All three factors, the wind shear, curvature created by an air parcel's position in a pressure pattern, and the curvature created by Earth's rotation (the Coriolis parameter), impose a spinning motion.

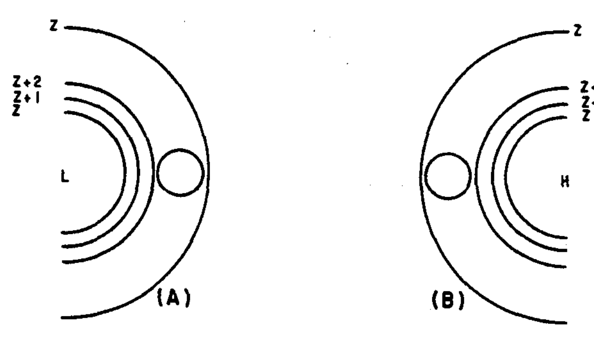

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Figure 8-5-3.—Illustration of shear effect opposing the curvature effect in producing vorticity. (A) Negative shear and positive curvature; (B) Positive shear and negative curvature.

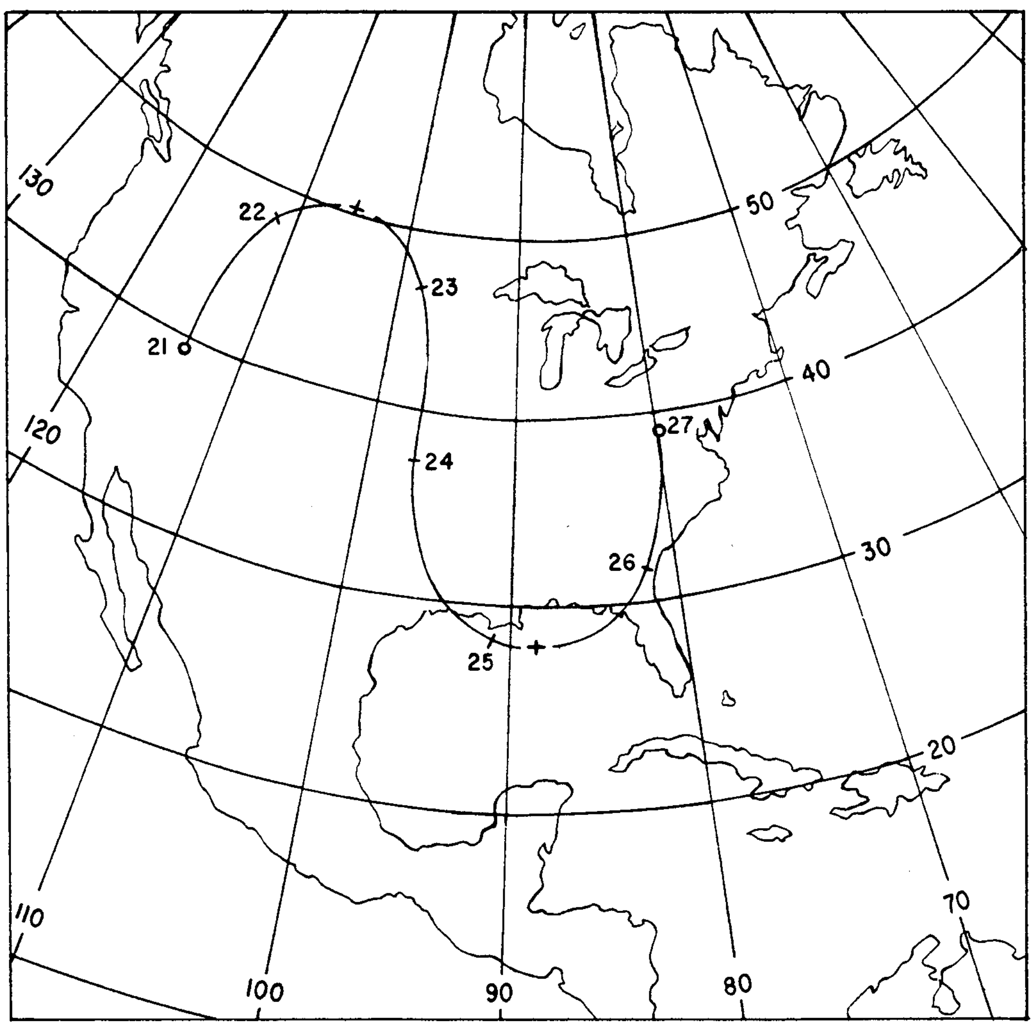
Therefore, absolute vorticity is equal to the Coriolis parameter plus relative vorticity.

Learning Objective: Define constant absolute vorticity trajectories and describe their use with upper-air charts.

CONSTANT ABSOLUTE VORTICITY TRAJECTORIES (CAVT)

It has been proved mathematically (under some rigid assumptions) that air parcels moving in the atmosphere conserve their absolute vorticity; that is, the absolute vorticity values do not change. The most important of the assumptions is that no convergence or divergence occurs. Actual observations under conditions of no convergence or divergence bear out this theory. Therefore, as a parcel moves from latitude to latitude in the Northern Hemisphere it describes a definite pattern called a constant absolute vorticity trajectory (CAVT). Figure 8-5-4 is an example of a CAVT. Constant absolute vorticity trajectory is defined as the latitudinal path of an air parcel whose absolute vorticity is unchanging.

Constant absolute vorticity trajectories are directly related to the movement of long waves.



SAMPLE C A V TRAJECTORY. (STROKES AND NUMERALS ALONG THE PATH MARK POSITIONS AND DATES FOR 24-HOUR INTERVALS FROM THE STARTING POINT ON THE 2300 EST MAP OF THE INITIAL DATE.)

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Figure 8-5-4.—Sample CAV trajectory.

The amplitude and wavelength of a long wave trough varies depending on the initial speed, direction, and latitude of the air parcel at the long wave's inflection points. Inflection points are simply those points on the long wave where wave curvature changes from cyclonic to anticyclonic and vice versa. These are shown in figure 8-5-5 as points A, B, C, and D.

The wind at the inflection points is the controlling factor in determining the future amplitude and wavelength of the long wave using CAVT tables. Both amplitude and wavelength increase with an increase in wind speed. Wind direction, however, may increase or decrease amplitude and wavelength, depending on the angle created by the wind direction at the inflection point (angle formed by a wind direction arrow and the latitude line). When the inflection angle is small, because of a basic westerly wind component, waves tend to be flat and elongated. On the other hand, a northerly or southerly wind component creates a large inflection angle, and waves exhibit great amplitude and wavelength.

Constant absolute vorticity trajectory tables were developed to provide a means of forecasting the future position of long waves. Because CAVT tables are used in forecasting, I will not elaborate

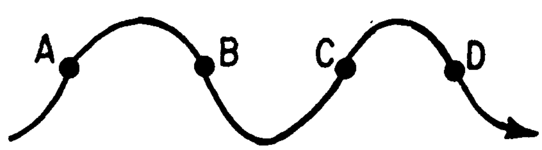

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Figure 8-5-5.—Sinusoidal vorticity path.

any further at this time. These tables will be discussed in more detail in the *AG 1* rate training manual.

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